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A Comparison of Present-Day and Pleistocene Proglacial Environments, with Particular Reference to geomorphology and Sedimentology.

Maizels, J. K

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A COMPARISON OF PRESENT-DAY AND PLEISTOCENE PROGLACIAL ENVIRONMENTS, WITH PARTICULAR REFERENCE TO GEOMORPHOLOGY AND SEDIMENTOLOGY

Part I

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Thesis submitted for the degree of
Doctor of Philosophy

University of London King's College

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ABSTRACT

Similarities and differences between present-day outwash deposits and their Pleistocene equivalents are demonstrated by reference to a proglacial valley-train in the French Alps and to four outwash deposits in the Vale of Strathmore and the southeast Grampians, Scotland. The major similarities between the deposits result from similar modes of formation, and include the logarithmic form of the longitudinal profiles, the braided channel patterns, the high width-depth ratios of proglacial channels, the large quantities of suspended and bedload sediments in meltwaters, and the similar facies types and depositional bedform structures. There are distinctive downstream changes in size, sorting, roundness and relative frequency of clasts of different lithologies, and these changes are compared with measurements in other proglacial deposits recorded in the literature. Different rates of downstream change are attributed to differential abrasion and sorting, to proglacial gradients and to glacier and meltwater conditions.

In contrast to their contemporary equivalents, the paleo-outwash deposits studied were affected by greater distances of transport, greater volumes of meltwater, more extensive post-depositional weathering, and possess a greater range of lithological types. The Scottish sediments are coarser, more acid, more massive and homogeneous than the Alpine valley-train sediments; the clasts are more rounded, and frequently pitted and weathered; postglacial dissection and isostatic readjustment have produced a series of tilted terrace deposits.

Local variability in horizontal and vertical size variations, estimated from detailed studies of a single bar and of a gravel pit section, results from changes in stream facies, in base-level, or in meltwater volume, the latter associated with the advance or retreat of the source glacier.

The proglacial environment in general is also discussed, with reference to the morphological development of the contemporary Alpine valley-train during a recent glacier advance, and to the proglacial microclimate and vegetation, meltwater régimes and groundwater flow.

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ADDITIONAL MATERIAL (in pocket)

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Chapter 1

I N T R O D U C T I O N

Chapter 1

I N T R O D U C T I O N

1.1. THE PROGLACIAL ENVIRONMENT

Beyond the margin of a glacier or ice-sheet lies a zone whose major geomorphological forms are influenced by the proximity of the ice. This zone is defined as the proglacial area. The microclimate, vegetation types, hydrology and morphological characteristics of the proglacial zone result from the activity of processes directly associated with an ice-mass situated upvalley. Many proglacial areas are characterized by extensive sedimentation by numerous meltwater streams, to produce fluvioglacial outwash deposits. These deposits may be confined between valley walls and are described as a valley-train, or they may extend beyond relatively narrow valleys to form a laterally extensive deposit, termed a sandur, an Icelandic term for a 'sand plain' (Spethmann, 1911).

Coarse fluvioglacial sediments accumulate most rapidly close to the ice-margin, so that both the gradient of the valley-floor and the size of sediments deposited gradually decrease as one proceeds down valley. The extensive deposition of outwash sediments is promoted by rapid and frequent changes in channel courses, often through lateral erosion of channel banks, resulting in the development of a braided network across the proglacial zone, and also by diurnal, seasonal and periodic variations in the discharge of heavily-laden meltwater streams, for each decrease in discharge is accompanied by deposition of sediment.

The combination of rapidly changing stream courses, of rapid rates of sedimentation and of frequent floods, creates an unstable ground-level environment, or substratum, for colonisation of outwash deposits by plants. Lack of vegetation, in its turn, delays the rate at which higher or temporarily 'abandoned' deposits may be stabilized by the

action of vegetative growth in slowing down flood currents which might otherwise scour the deposit. Plant colonisation is also delayed by a fairly hostile micro-climate, in which maximum temperatures are consistently cooled by cold air currents blowing off the ice mass.

Recent studies of proglacial environments have mostly been concentrated in the higher latitude areas including the sandar of Iceland (Hjulström, 1952, 1954a,b, 1955; Sundborg, 1954; Krigström, 1962; John and Sugden, 1962; Price, 1969, 1971; Jewtuchowicz, 1971, 1973; Bogacki, 1973; Churski, 1973; Galon, 1973a,b; Galon and Szupryczynski, 1973; Klimek, 1973; Konysz, 1973; Kozarski and Szupryczynski, 1973; Wojcik, 1973a,b), Alaska (Viereck, 1966; Ehrlich and Davies, 1968; Slatt and Hoskin, 1968; Fahnestock, 1969; Broscoe, 1972; Boothroyd, 1972), Spitsbergen (Jewtuchowicz, 1962), and Baffin Island (King and Buckley, 1969; Church, 1970, 1972). Valley-trains frequently occur in front of more temperate glaciers, such as those in the Rockies (Fahnestock, 1963; McDonald and Banerjee, 1969, 1971), the European Alps (Vivian, 1967a; Maizels, 1970, 1973), Norway (King, 1959), the Alps of New Zealand (Speight, 1940) and the Himalayas (Hewitt, 1967, 1969). The latter areas seem to have attracted less attention, partly because the valley-trains are on a smaller scale than sandur-plains, and partly because it is thought that the environment may not be strictly comparable with Pleistocene outwash environments.

1.2. PREVIOUS STUDIES OF PROGLACIAL OUTWASH DEPOSITS

1.2.1. EARLY OBSERVATIONS

As early as 1841, Charpentier described crudely stratified deposits, incorporating some rounded particles, at the foot of most Alpine glaciers. He also noted that plants began to colonize these deposits during the meltseason. During the late nineteenth century and the early years of the twentieth, the existence of extensive proglacial meltwater deposits was reported by a number of workers, not only in the Alps, but in Iceland, Greenland and

North America. In Alpine regions glacial geomorphological research was dominated by studies of the glaciers themselves (e.g. Agassiz, 1840; Viollett-le-Duc, 1877; and later Rabot, 1902,a,b). Only a few workers mentioned the characteristics of, or processes operating in, the proglacial environment. However, in 1902, Girardin gave a fairly detailed account of the "laisses glaciaires" of the Western Alps. His observations included changes in roundness, size and sorting of pebbles with increasing distance from the glacier, the high quantities of suspended sediments in the meltwater streams, and also the approximate rates at which the development of humus and vegetation growth took place on proglacial deposits. Pleistocene outwash terraces, or 'Deckenschotter', in the Alpine region formed the basis of the glacial chronology worked out by Penck and Brückner (1909). The relative age of terrace deposits was determined by the degree of weathering, the development of a soil profile, stratigraphy, morphology, and the terminal moraine with which a terrace was associated.

Studies in North America were completed by Chamberlin (1883) and Salisbury (1892) who concentrated on the investigation of Quaternary valley-trains and outwash (or "overwash") plains, and they developed the basis of the North American Pleistocene glacial sequence. Subsequent expeditions to Alaska by Gilbert (1904), Tarr (1909) and Martin (Tarr and Martin, 1914) involved observations and descriptions of some contemporary proglacial processes and bedded sediments of the "gravel plains" and of downvalley gradations of particle sizes. However, emphasis tended to lie on the study of glacial processes and landforms rather than on those of the proglacial environment. For several decades investigations were directed increasingly towards glacier studies, particularly in relation to factors affecting the mass budget of a glacier, and consequent historical variations in the position of the ice-snout (e.g. Ahlmann, 1922, 1927, 1953; Faegri, 1948; Thorarinsson, 1940). Only recently has the detailed study of proglacial outwash deposits attracted research workers, and they tend to be associated with separate "schools" of investigators.

1.2.2. STUDIES IN 'GLACIAL GEOMORPHOLOGY'

A major aim of glacial geomorphologists has been to develop techniques for differentiating between fluvioglacial and glacial deposits, deposits of different ages and of varied origin ('kames', 'eskers', 'proglacial outwash' etc.). The methods employed to aid identification of the environment of deposition include observations or measurements of the particle lithology and shape (e.g. Hövermann and Poser, 1951; Galloway, 1956; Cailleux and Michel, 1967; King and Buckley, 1968), the degree of weathering, the morphology and the stratigraphic position characterizing a particular deposit. These methods have been applied to both present-day and Pleistocene fluvio-glacial deposits.

Important contributions to the literature on the geomorphology of present-day bodies of outwash include those by Jewtuchowicz (1955, 1971, 1972, 1973). He suggested that the chief elements of an outwash deposit which should be studied in order to allow objective comparisons with paleodeposits of similar origin should include the texture, shape, size, sorting and orientation of the constituent particles and their relation to the morphology and stratigraphical units of the whole deposit. His interpretation of the depositional environment of a Polish Pleistocene outwash body was based on field work in Iceland and Spitsbergen. Price (1965, 1969, 1971, 1972, 1973) has also made some major contributions towards elucidating recent temporal and spatial variations of geomorphological features, both erosional and depositional, at the margins of Icelandic and Alaskan glaciers and ice-caps. Other workers on the geomorphology of proglacial areas include Sundborg (1954), Jayet (1955), King (1959), John and Sugden (1962), Ashwell (1966), Vivian (1967a), Evans (1969) and Broscoe (1972). These, and many other workers, have carried out detailed and small-scale studies of particular ice-marginal and proglacial areas, and have made useful contributions to understanding the variability of geomorphological features which may exist close to an ice-margin.

The identification and description of Pleistocene outwash deposits have been carried out in almost all areas which were glaciated during that period. One of the main aims of these studies has been to determine the agent and environment of deposition. A second aim has been to establish a regional glacial chronology which might then be correlated with well-established chronologies elsewhere. Often, however, relative dates of formation are assigned not directly to the outwash deposits, but to the glacial tills with which they appear to be associated. Recent work along these lines which also emphasizes the character of, and possible processes forming the outwash deposits, includes Frödin (1954) and Hoppe (1959) in Scandinavia, Cailleux (1960a)and Seret (1965, 1968) in northwest continental Europe, Kirby (1969a)and Cullingford and Smith (1974) in Scotland, Horberg (1954), Denton and Stuiver (1967), and Porter and Denton (1967), in North America, Church (1970) in Baffin Island, and Suggate (1958, 1963), Gage (1958) and Speight (1963) in New Zealand. Summaries of some of the major works are contained in Flint (1971), Charlesworth (1957), Tricart and Cailleux (1962), Embleton and King (1968), Davies (1969), Tricart (1970), and Price (1973).

Glacial geomorphologists have, however, tended to avoid detailed analyses of the processes and the nature of the sediments (e.g. structures, size distributions) that result. They have rarely realized that, through such detailed study of outwash deposits, valuable additional information may be gained about the depositional environment. The types of information which might be extracted are outlined below:

1) Hydrodynamic conditions of deposition such as meltwater velocity, depth of stream flow and flow regime;

2) The source, or sources, of the components of the deposit, the route of transport and the effects of fluvial transport;

3) The rates of erosion or aggradation of the proglacial sediments;

4) Explanation of the variability of sedimentary facies, including sedimentary structures, lithology, shape and texture of the deposit, both within a particular deposit and between different proglacial outwash deposits.

1.2.3. STUDIES IN 'FLUVIAL GEOMORPHOLOGY' AND 'HYDROLOGY'

Present-day proglacial environments have also attracted those geomorphologists interested in the action and products of fluvial processes. In the proglacial environment fluvial processes are (1) speeded up, as a result of the rapid and large changes in the discharge, sediment content, shape and position of the braided meltwater streams; and (2) on a sufficiently small scale to be observed in great detail over downstream distances, over a long time-period, and in a variety of hydrodynamic situations. In addition, observations are frequently unhindered by dense vegetation, steep slopes or by human or animal interference, compared with more temperate fluvial basins.

Detailed study of the morphology and hydrology of proglacial outwash areas was initiated on a large-scale by Hjulström and his colleagues on the expeditions to the Hoffellssandur in Iceland in 1951 and 1952 (see also Thorarinsson, 1939). The Scandinavian workers produced a series of important papers which provided valuable additions to the summary of information on processes of outwash formation and the methods of deposition in a braided fluvial environment. Several important theoretical conclusions were developed from the data collected from the proglacial streams. Hjulström (1935), for example, had derived an erosion-curve graph, in which he calculated the range of stream velocities required to initiate and to terminate the movement of particles of different sizes on the stream bed. In 1955, Hjulström attempted to calculate

the rate of ground-water flow as a percentage of total meltwater discharge from the Hoffellssandur, and concluded that it represented only 0.4 per cent of the total. The Hoffellssandur also provided Hjulström with observations from which he could explain the probable methods of development of a braided channel network (e.g. Hjulström, 1952, 1954). Arnborg (1955) related meltwater discharge amounts to climatic conditions and the rate of ablation of the source glacier. He found that discharge was also determined by the drainage of ice-dammed lakes and the amount of surface run-off in the basin. His conclusions were based on measurements of meltwater discharge throughout the meltseason, on the proglacial River Austerfljot.

More recently, proglacial hydrology has been intensively studied by such workers as Krigström (1962) and Stenborg (1965, 1970). Krigström described the sedimentary processes operating in the braided fluvial system, and calculated the minimum meltwater discharge required to submerge the whole valley-train surface. He concluded that discharge would have to be so large that the outwash surface is unlikely to be completely flooded at any time. Stenborg related meltwater flow to conditions within the source glacier, and identified the different sources of water which contribute to winter meltwater discharges (see Chapter 2). The large volumes of suspended sediments and of minerals in solution have attracted other workers, such as Lanser (1958), Borland (1961), Keller and Reesman (1963), Østrem, Bridge and Rannie (1967), Liestøl (1967), Vivian (1970c), Church (1972), Churski (1973), Klimek (1973), Miller and Helmers (1973), Pinchak (1973), and Vivian and Zumstein (1973). Their studies have included measurements of the relationship between meltwater discharge and suspended sediment content, which was found to be particularly dependent on the rate of discharge change and the temperature of the water.

1.2.4. STUDIES IN 'SEDIMENTOLOGY'

Sedimentologists have been working in a sphere of research closely associated with that of the fluvial geomorphologists and hydrologists. However, their emphasis lies in determining the relationships between stream process and the resultant sedimentary structures and textures. Their results have proved fairly successful in that they are applicable to a wide range of water-laid sediments, and data obtained from laboratory experiments generally correspond closely with the results of field measurements. Many of the verifications of laboratory investigations, which have been obtained in present-day proglacial streams, are still continuing.

Doeglas (1962) gave a description of the chief characteristics of the hydrological and sedimentary facies of a proglacial braided environment, relating variations in meltwater discharge to variations in the types of sedimentary structures. A detailed analysis of the relationships between sediment type, meltwater discharge variations and channel morphology was carried out by Fahnestock (1963) in the White River (Washington State). He also discussed the wide range of processes operating on the valley-train, the contribution to valley-train sediments by till and mudflows, the factors contributing to the development of a braided channel pattern, and the short-term changes in this pattern, measured in days and hours.

Other workers have concentrated only on sediment characteristics. Ehrlich and Davies (1968), working in front of the Mendenhall Glacier, Alaska, attempted to relate relative rates of downstream decrease in the proportions of different particle lithologies to distance from source and to particle resistance. Slatt and Hoskin (1968) investigated similar problems in front of another Alaskan glacier, the Norris Glacier, but they concentrated on the characteristics of the fine sediments, including heavy mineral suites and textures of the fine fractions.

They related the percentage of quartz in the sediment to the degree of erosion and transport within the source glacier, and to the intensity of weathering.

Studies of the relationships between meltwater discharge, sedimentary structures and depositional facies on the one hand, and valley-train or sandur morphology on the other hand, have proved increasingly important recently, especially in North America (e.g. Fahnestock, 1969; Williams and Rust, 1969; McDonald and Banerjee, 1969, 1971; Rust, 1972; Church, 1972; Bradley, Fahnestock and Rowekamp, 1972; and Boothroyd, 1972). Recent work by other North American sedimentologists has concentrated on the analysis of a variety of sedimentary structures within examples of contemporary and Pleistocene outwash deposits (e.g. Jopling and Walker, 1968; Church, 1970, 1972; Costello and Walker, 1972; Stanley and Wayne, 1972; Eynon and Walker, 1974; Spoljaric, 1974). These workers have aimed to contribute to the development of a model of the 'braided fluvial environment'. The model construction was based on two approaches. Firstly, they aimed to determine stream processes, such as flow régime and rates of sediment transport, by detailed study of the sedimentary structures. Interpretation was possible through comparison with empirical and laboratory investigations. Secondly, they attempted to gather information on the degree of sediment variability over time at a single locality as expressed by vertical variations in sections excavated in paleo-deposits. The latter analyses were to provide a basis for recognizing any cyclic or systematic pattern in the type of sedimentation in the braided channel environment.

1.2.5. DISCUSSION

Proglacial outwash environments have thus served as field study areas for the development of models of the braided fluvial environment by both fluvial geomorphologists and sedimentologists. Furthermore, many of the hydrologically-oriented investigations are closely aligned

with developments elsewhere in fluvial geomorphology. Hydrological studies in proglacial areas have contributed to the discussions on rates of sediment movement of heterogeneous bedloads and on channel shape (e.g. Sundborg, 1956), on the development of a 'graded', longitudinal profile and of a fluvial system in equilibrium (e.g. Church, 1972), and on the relations between fluvial process and the resulting sedimentary structure (e.g. Rust, 1972; see also Rana, Simons and Mahmood, 1973; Culling, 1957a,1960; Hack, 1960; Simons, Richardson and Nordin, 1965; Leopold and Wolman, 1957).

The fact that the deposits are so closely associated with a contiguous glacial source appears to have been ignored in many of the studies under consideration. A few workers (e.g. Fahnestock, 1963; Church, 1972) have discussed the possible effects of deglaciation on proglacial stream discharge and sediment transport, rates of aggradation or erosion, evolution of the longitudinal profile, and, more recently (e.g. Costello and Walker, 1972) on the stratigraphy. An ice mass has otherwise been simply recorded as the source of debris and water, and further examination has been considered unnecessary. It is unusual, therefore, to find any analytical reference to the nearby glacier, although cyclic and periodic variations in meltwater and debris release from the glacier may be major causes of the large internal variability observed in the nature of the sediments composing proglacial outwash deposits. An important task that remains is to determine the probable causes of this variability both within a single deposit, and also between different outwash bodies, in terms of relative amounts of meltwater and debris release from the ice source.

1.3. AIMS OF THE PRESENT STUDY

This study represents an approach to the interpretation of paleo-outwash sediments in the light of observations in a present-day proglacial environment. Two study areas were selected; the first, a contemporary Alpine valley-train of the Bossons Glacier, in the French Alps; the second, a Late Pleistocene series of outwash deposits in northeast Strathmore, Scotland (see section 1.4.). In order to reach a stage at which sufficient information was available from the present-day environment for valid comparison with that from paleo-deposits, the project was structured to pursue four objectives. These are briefly outlined below.

(1) To describe (a) the morphological features of a present-day proglacial environment and (b) some of the major processes which can operate in different parts of a proglacial environment at different times, to produce the observed morphological features.

(2) To determine the effects that morphology and fluvioglacial processes may have on variations in the characteristics of the sediments composing an outwash deposit.

(3) To use the results of (2) to interpret some of the sediment characteristics of the paleo-outwash deposits in terms of the possible processes producing them and the morphological conditions at the time of deposition.

(4) To account for any differences between the deposits of the active proglacial area and those of the paleo-proglacial areas in terms of

(a) Nature and strength of processes operating;

(b) Regional environmental conditions, or the "boundary conditions" of Krumbein and Sloss (1968). This may include nature of bedrock lithologies, distance of particle transport from the ice-margin, and possibly the nature of the glacier regime; and

(c) Post-depositional alterations, both chemical and mechanical in origin.

1.4. CHOICE OF FIELD STUDY AREAS

1.4.1. CONTEMPORARY ALPINE VALLEY-TRAIN

The valley-train of the Bossons glacier, situated on the north-facing slope of the Mont Blanc massif, was selected as the field area for study of a present-day proglacial environment (figure 1.1, page 43). The valley-train has several advantages for this type of study:-

(1) It forms a compact and well-defined deposit;
(2) Analysis of sediment characteristics is eased by the fact that there are only two major lithological types present, namely (a) micaceous and calcareous schists, which form the dominant lithological group, and (b) 'protogine', or gneissose, granites.
(3) The history of glacier movements, both of the Bossons glacier and of other glaciers in the massif, is well-documented. Climatic data and records of meltwater discharge for locations close to Bossons, are also available for study.
(4) The ice-snout and proglacial valley are readily accessible even in early spring, so that observations could also be made before the meltseason had commenced.

1.4.2. PLEISTOCENE OUTWASH DEPOSITS OF NORTHEAST STRATHMORE, SCOTLAND

The selection of a set of Pleistocene outwash deposits which would allow valid comparisons to be made with the Bossons deposits was a major problem. Four outwash series were finally selected, of which one was studied in greater detail than the other three. They are located in the County of Angus in the northeast of the Vale of Strathmore, having been laid down by meltwaters emerging from glaciers in the following present-day river basins of the southeast Grampians (figure 1.2, page 44):

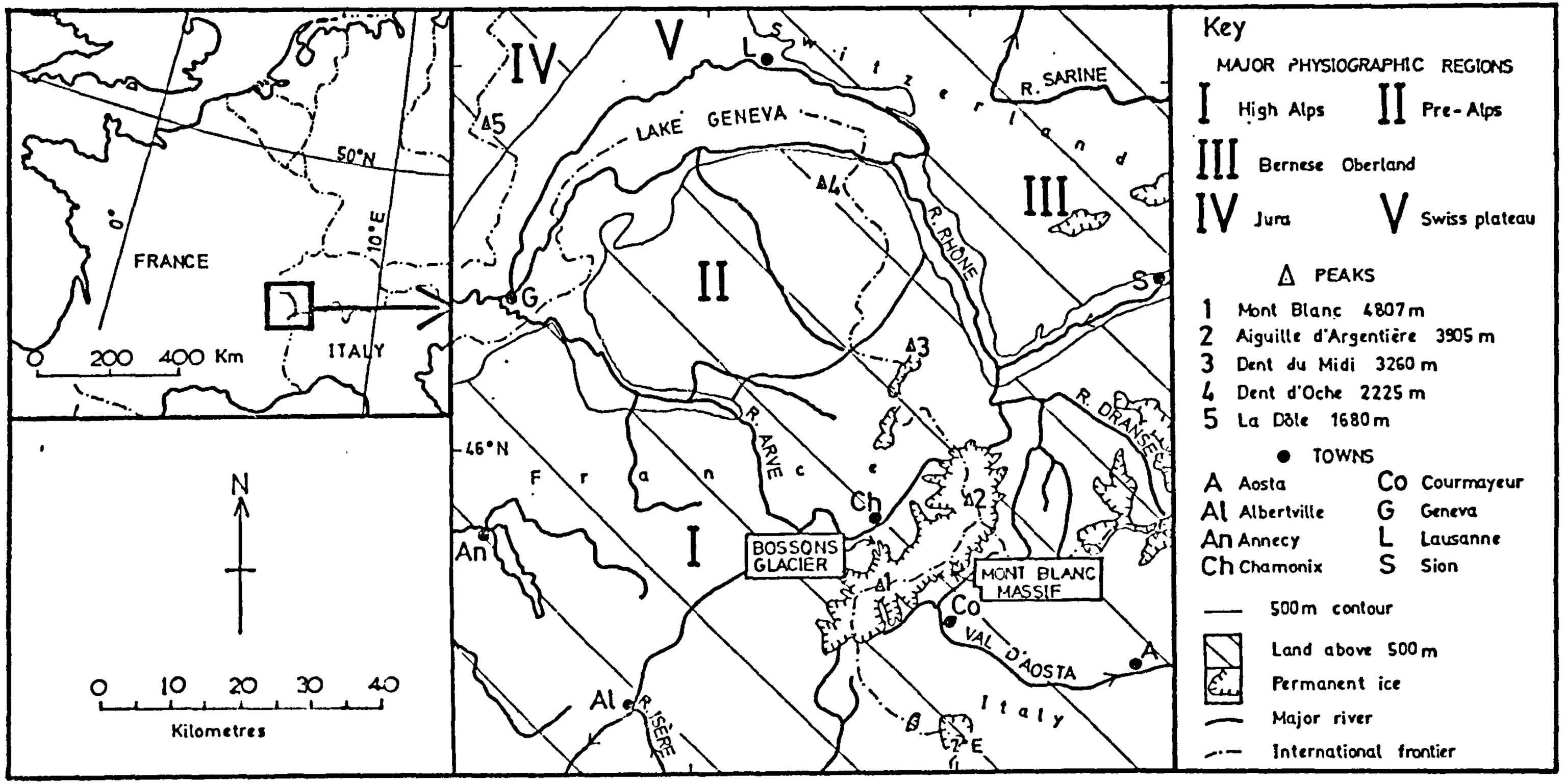


Figure 1.1. THE LOCATION OF THE MONT BLANC MASSIF AND THE BOSSONS GLACIER.

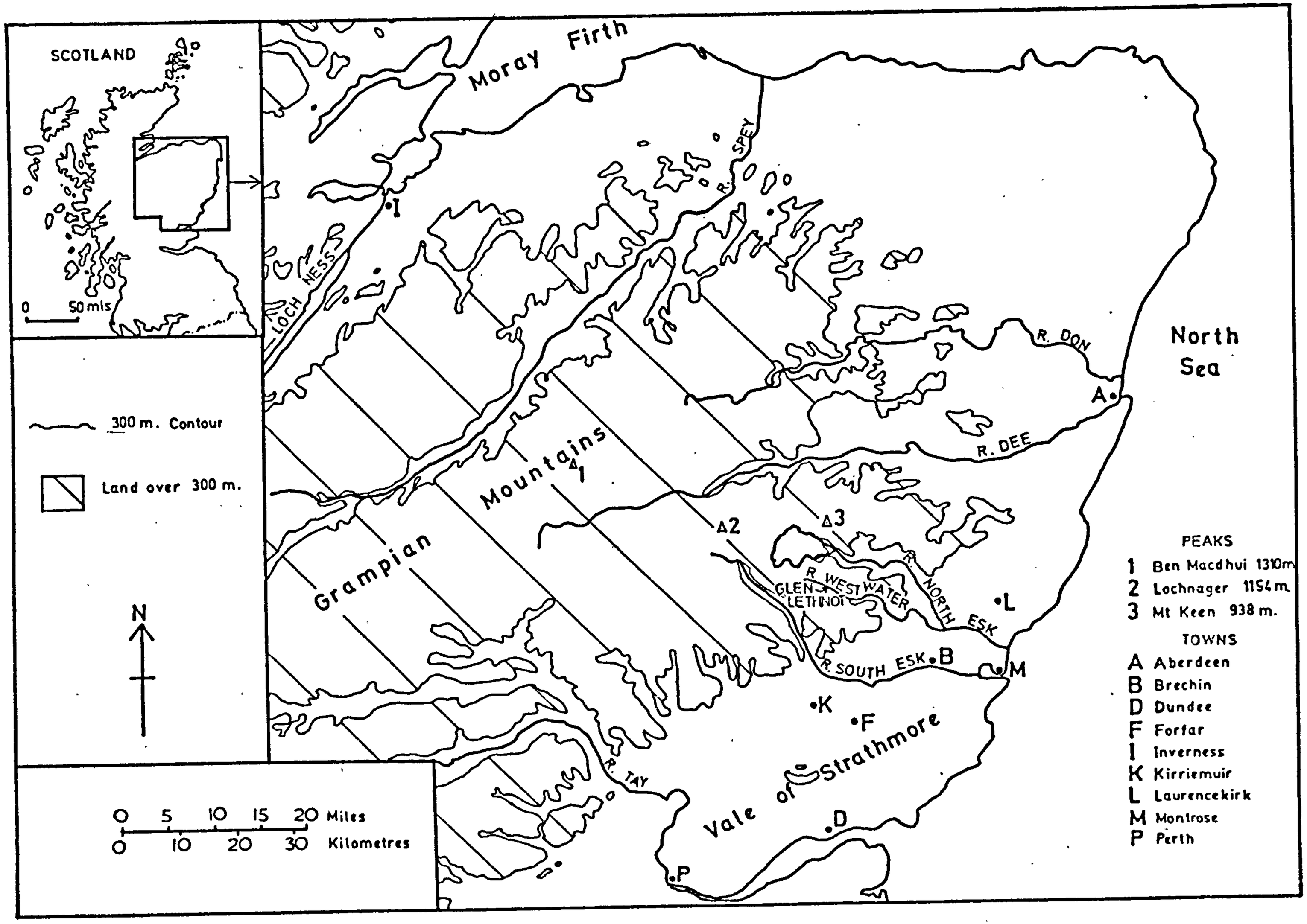


Figure 1.2. THE LOCATION OF THE FOUR SCOTTISH RIVER BASINS

(1) the River North Esk, and
(2) its major tributary, the River Westwater;
(3) the River South Esk, and
(4) Glen Lethnot, the headwater valley of the River Westwater.

These outwash deposits were selected for three reasons:

(1) It was initially considered that they were relatively well-dated, although recent research (e.g. Paterson, 1974) suggests some uncertainty about the precise data of formation. Nevertheless, there is no doubt that they were all deposited during Late Weichselian times, the last glacial phase of the Pleistocene.
(2) The deposits were laid down by meltwaters derived from valley glaciers whose basins comprised lithologies similar to those at Bossons, namely, schists and granites. The major deposits (North and South Esk, and River Westwater) also included six other lithological groups, but in all deposits studied, schist was the major lithology present.
(3) These outwash deposits contain a relatively large number of exposures, including several gravel pit excavations, allowing detailed, localized analyses to be carried out.

The River North Esk outwash deposits are particularly well exposed in comparison with the other series, and for this reason was the first region to be selected, and was studied in greater detail. The River Westwater was found to have been a major contributory source of sediment to the North Esk outwash series, and it was therefore also included in the study.

The River South Esk outwash deposits are more extensive, although less well exposed, than the North Esk deposits. However, it was felt desirable to gain information from an additional paleo-deposit for, while there is already a good deal of literature to support the studies on the present-day fluvioglacial deposits, there is a paucity of comparative paleo-outwash studies.

The morphology of these three outwash deposits approaches a sandur-like deposit rather than that of a valley-train. Field-work was therefore extended to the shorter headwater valley of the River Westwater, named Glen Lethnot, where terraces, apparently dissected valley-train deposits (Sissons, 1972), occupy the valley-floor.

1.4.3. DISCUSSION

Difficulties are likely to arise when attempting sound comparisons of deposits in areas in which there are differences in scale, exposures, gradient, age, geology, glacier regime and relative position of the ice-margin. These factors are considered in some detail in Chapter 4. The main problem seems to be whether or not relationships between sediment and process characteristics of contemporary fluvioglacial deposits are applicable to fluvioglacial deposits which

(1) were formed much earlier by meltwaters derived from larger, colder glaciers, and which
(2) are observed from isolated, vertical sections, rather than a total exposed surface.

Despite these differences relating to local conditions, it must be emphasized that all these deposits originated as a result of the deposition by proglacial meltwaters, of sediments derived largely from within the glaciated area, carried initially within an ice mass, and subsequently by streams extending beyond the ice-margin. It is therefore likely that similar processes have produced similar types of end-product deposits. The present study sets out to identify and account for the similarities and differences in the morphology and sediments of the contemporary and the paleo-outwash deposits.

1.5. PLAN OF THE STUDY

The thesis begins with a local, regional study of the Alpine proglacial environment; Chapter 2 ('Physical background') deals with factors which affect the geomorphology of the Bossons proglacial environment, and Chapter 3 discusses the morphology of this proglacial outwash body. The regional background to the Strathmore deposits is discussed in Chapter 4 and some of the problems involved in comparing the two series of deposits are introduced. The results of these comparisons are set out in the following three chapters, where emphasis is placed on meltwater and channel conditions and on sediment characteristics. The final chapter summarizes some of the more important relationships between fluvioglacial processes, outwash morphology, the distribution of sediment types, and other physical aspects of the proglacial environment.

This study has emphasized the development of working hypotheses and ideas to be followed up in future research, rather than the presentation of a series of firm conclusions. Much work remains to be done, especially in the sphere of hydrological processes, but it is hoped that some advance towards more complete understanding of some of the features characterizing proglacial sediments has been accomplished.

Chapter 2

B O S S O N S I:

T H E P H Y S I C A L B A C K G R O U N D

Units of measurement

The following metric units are employed in the text:

km	kilometre
m	metre
cm	centimetre
mm	millimetre
kg	kilogramme
g	gramme
mg	milligramme
l	litre
ml	millilitre
yr	year
s	second
^{o}C	degrees centigrade

Conventional abbreviations

A.D.	years after Anno Domini
B.P.	years Before Present (i.e. 1950)
O.D.	height above Ordnance Datum
p.a.	per annum
ppm.	parts per million
$\log_n$	logarithm to the base n

List of symbols

Some symbols have several meanings because of an overlap in standard usages; the meaning of most symbols is further clarified in the text, together with units of measurement. Additional lists of symbols are provided with each relevant chapter.

A	glaciated area
b	'beta-coefficient', or slope coefficient in regression equation
Q	meltwater discharge

Chapter 2

BOSSONS I: THE PHYSICAL BACKGROUND

This chapter deals with the physical background to the Bossons glacier and its proglacial valley. Studies by the author complement the large amount of published information on the geology, relief, climate, glacial history, hydrology and biogeography of the Mont Blanc massif and the Bossons glacier basin.

2.1. GEOLOGY

2.1.1. THE MONT BLANC MASSIF

The Mont Blanc massif extends over an area of about 1200 square kilometres. It represents an acid igneous intrusion, forced upwards through pre-Carboniferous altered sediments, an upward movement which culminated during the Pontian-Lower Pliocene period, and known generally as the 'Alpine Orogeny'. The Mont Blanc massif trends northeast-southwest, and forms one massif within the chain of 'external crystalline massifs'. These represent the areas to which, and over which, the Alpine nappes were thrust from their root, or 'internal', zones, situated to the southeast (figure 2.1, page 51).

The Mont Blanc massif is also closely associated with the neighbouring massif of the Aiguilles Rouges which lie to the northwest of the Mont Blanc massif, and are separated from it by a narrow tectonic depression known as the Chamonix syncline. The syncline is composed of Mesozoic and Cainozoic sediments, overlain by recent outwash sediments deposited by meltwaters derived from the five major glaciers situated on the northwest slopes of the Mont Blanc massif.

The massif itself is composed of two major petrological groups.

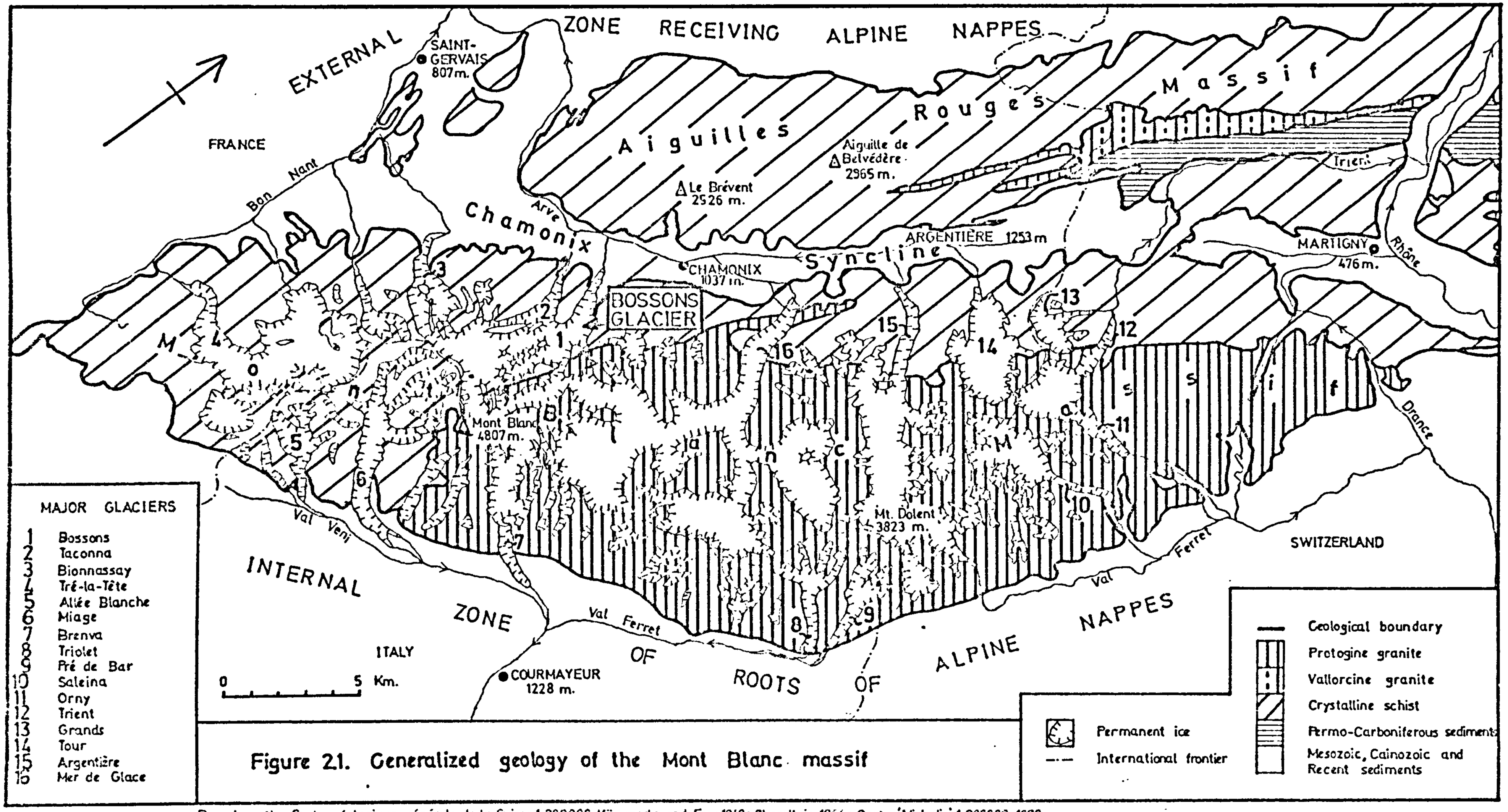


Figure 2.1. Generalized geology of the Mont Blanc massif

Based on the Carte géologique générale de la Suisse, 1:200,000, Künmerly and Frey, 1948; Charollais, 1966; Carte 'Michelin', 1.200000, 1970

(1) The core of the massif is formed of 'protogine' granite, which is of early Carboniferous age (Bonney, 1912). The original composition and structure of the protogine granite were later affected by Alpine earth movements (Tyrrell, 1926; Lombard, 1955). The main minerals, biotite and orthoclase feldspar, have, for example, been altered respectively, to chlorite and to sericite or kaolinite (see also Duparc and Mrazec, 1898; Abrard, 1948; Collet, 1948). The small amount of free quartz present has remained relatively unaffected. The original granite structure has been partially altered to produce what Bonney (1912) describes as a "granitoid gneiss" (or "gneiss oeillé"), for parts of it exhibit a rough alignment of minerals and a coarse, schistose foliation.

(2) A metamorphic aureole, which is largely composed of crystalline schists, surrounds the protogine core. The schists are derived from the metamorphism of pre-Carboniferous sediments and Precambrian schists (Charollais, 1966). Most of the schists are rich in muscovite, biotite and feldspar, especially plagioclase feldspars which may be sericitized. The schists also include grains of quartz and epidote. Some varieties, such as the "sericitic gneiss", are relatively coarse-grained, so that the crystalline rocks form a continuous series, ranging from coarse-grained granite through "granitoid gneiss" and "sericitic gneiss" to crystalline schist (Gignoux and Moret, 1952; Moret, 1962).

2.1.2. THE BOSSONS VALLEY

Description of the distribution of bedrock types in the Bossons valley is based on (1) the geological map published by Corbin and Oulianoff in 1927-1928, at a scale of 1:20 000, and (2) a more recent, smaller-scale geological map published by the Commission Géologique Suisse in 1948, at a scale of 1:200 000. Each map series is accompanied by an explanatory geological guide (Notice explicative de la carte géologique du massif du Mont Blanc, 1927 and 1928; Collet, 1948; Lombard, 1955). However, some of the bedrock

contacts in the Bossons valley are masked by a permanent ice cover.

The Bossons glacier originates in two névés: a western and an eastern névé (figure 2.2, page 54) which are situated in different bedrock zones. The eastern névé lies exclusively within the protogine granite zone. The western névé, by contrast, lies mainly within the zone of crystalline schists, and only in the extreme southeast of this névé field does ice overlie granitic bedrock. The névés converge below an altitude of about 3000 m to form the glacier, which then flows entirely on schist bedrock for a distance of about eight kilometres to its terminus. The debris transported downvalley by the glacier therefore includes mainly granite on the east, and schist, with smaller quantities of granite, on the west, a distribution which is reflected as far as the proglacial valley sediments (chapter 7.3).

2.2. RELIEF

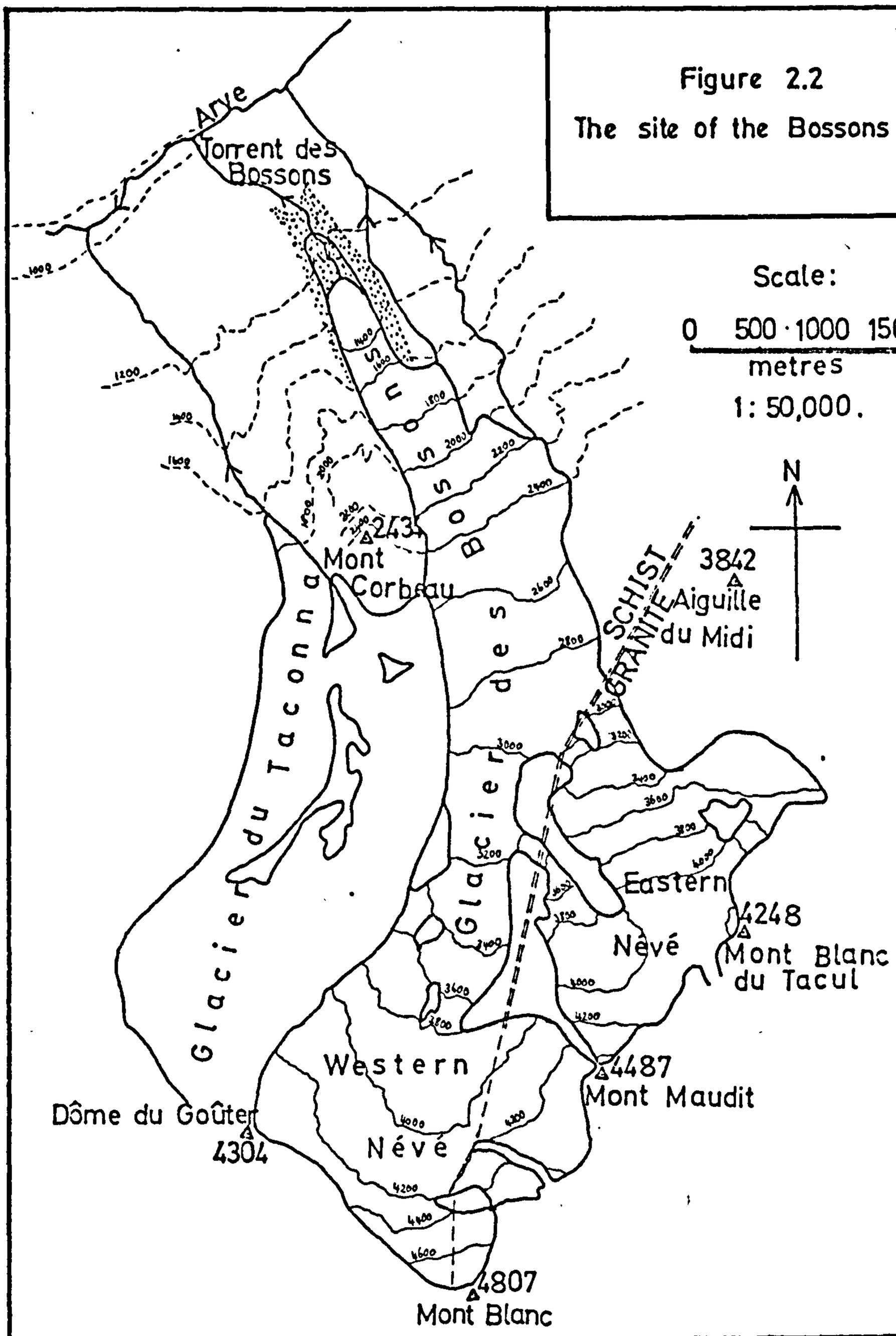
2.2.1. THE MONT BLANC MASSIF

Mont Blanc rises to 4807 m and is situated in the southern section of the massif whose crestline is formed of a series of major peaks, each higher than 3400 m. Steep gradients characterize the massif as it is only about 15 km wide and yet rises up to 3000 m above the surrounding valley floors. The relief of the massif is slightly asymmetrical in that the main watershed lies closer to the southeast than to the northwest boundary. Slopes on the southeast, Italian flank therefore tend to be much steeper than those on the northeast, French side. The mean gradient between Mt. Dolent and the Val Ferret is 1:1.9 (or $27^{o}41'$) compared with a mean gradient of 1:4.48 (or $12^{o}34'$) between Mt. Dolent and Argentière (see figure 2.1).

The massif has been dissected into numerous cirque regions lying below the main crest ridge. These cirques are the nourishing grounds for the massif's 16 major glaciers, which flow radially outwards towards the bounding valleys (see figure 2.2).

Figure 2.2

The site of the Bossons glacier



KEY

- Ice contours on Bossons glacier, at intervals of 200m.
- Ground contours below 2400m, at intervals of 200m.
- 4304 — Summit height in m.
- Stream
- Moraine associated with Bossons glacier
- Schist - Granite boundary

Based on Maizels, 1970

2.2.2. THE BOSSONS VALLEY

The basin of the Torrent des Bossons covers 13.98 km^2, of which 75.11 per cent was ice-covered in 1968 (as measured on the I.G.N. 1: 10 000 Topographic map). The Bossons glacier corresponds to a "Valley Glacier Type II" according to Ahlmann's classification (Baussart, Carbonnell and Cazabat, 1965), for fifty per cent of the area of the Bossons glacier lies between 3560 m and 4800 m, i.e. within the cirque areas, and seventy per cent lies above the 'equilibrium line', which is situated at 2800 m according to Corbel (1963) (compared with 3100 m according to Paschinger in 1911). The remaining 30 per cent of the glacier area lies below 2800 m, and the glacier terminates at an altitude of approximately 1230 m.

A low-gradient valley-train extends for over 200 m beyond the ice-snout, overlooked on both sides by steep, abandoned lateral moraines, rising about 100 m above the valley floor. The downstream end of the valley-train is marked by a sudden break in slope, from a gradient of less than 1^{o} to one of over 30^{o}. The area of braided channels becomes one drained by a single, steeply graded channel interrupted by a series of rapids. This is the Torrent des Bossons. The stream continues over a 'fluvioglacial fan' (Veyret, 1969) to the floor of the Chamonix valley, where it joins the River Arve, a tributary of the River Rhone, northwest of the village of Les Bossons, at an altitude of about 1000 m (figure 2.2.).

The longitudinal profile of the glacier and its meltwater stream, the Torrent des Bossons, is marked by several distinctive features (figure 2.3, page 56).

(1) Ice moving from the western névé travels 8 km to the ice-snout, while ice from the eastern névé travels only 6 km. Other factors being equal, the greater opportunity for bedrock erosion and particle breakage over the longer distance may have some measurable effect on the size of debris reaching the ice-snout. The distribution of granite particles on the east and schist particles on the west of

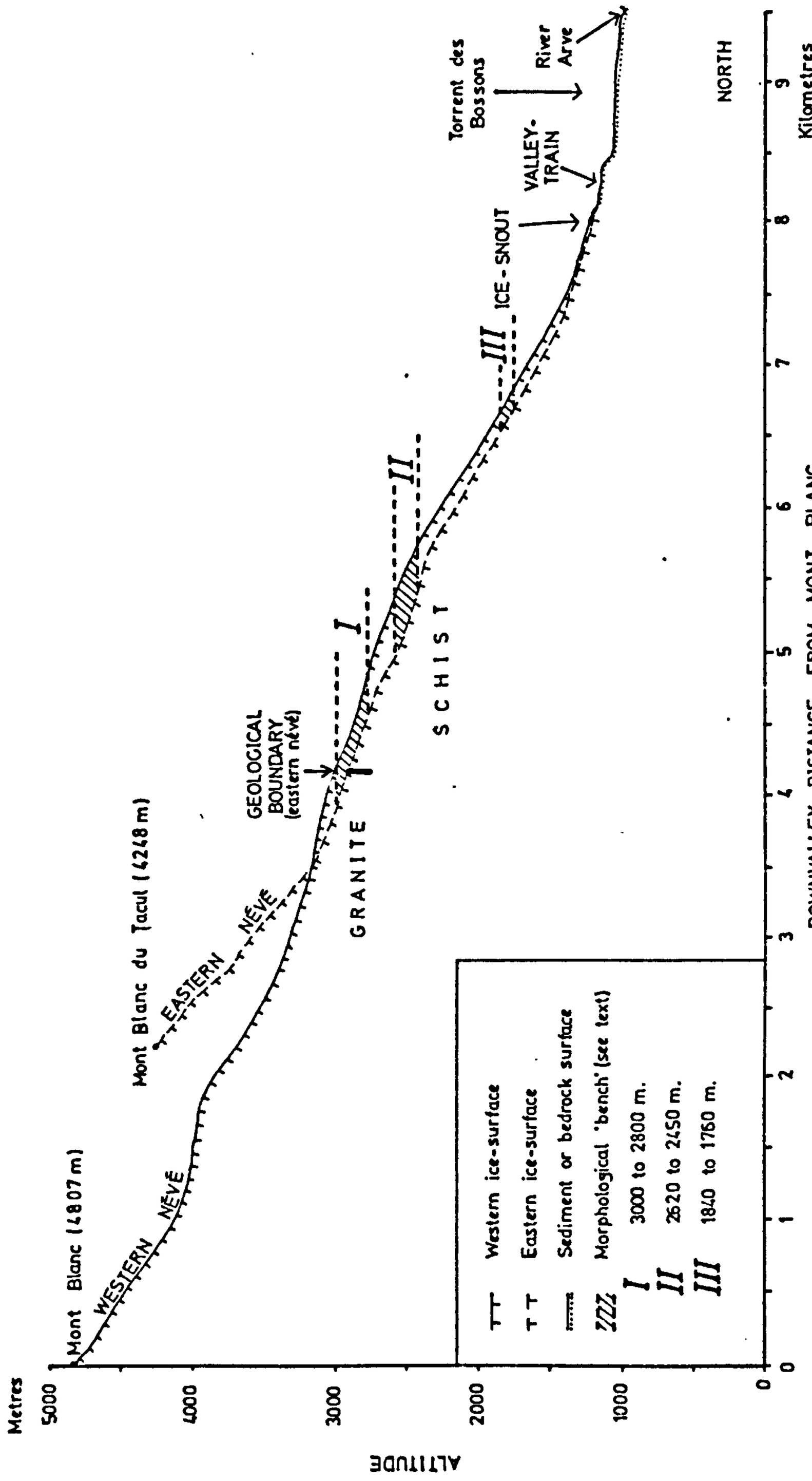


Figure 2.3. The long-profile of the ice-surface of the Bossons glacier

the proglacial valley indicates that little transverse ice flow takes place.

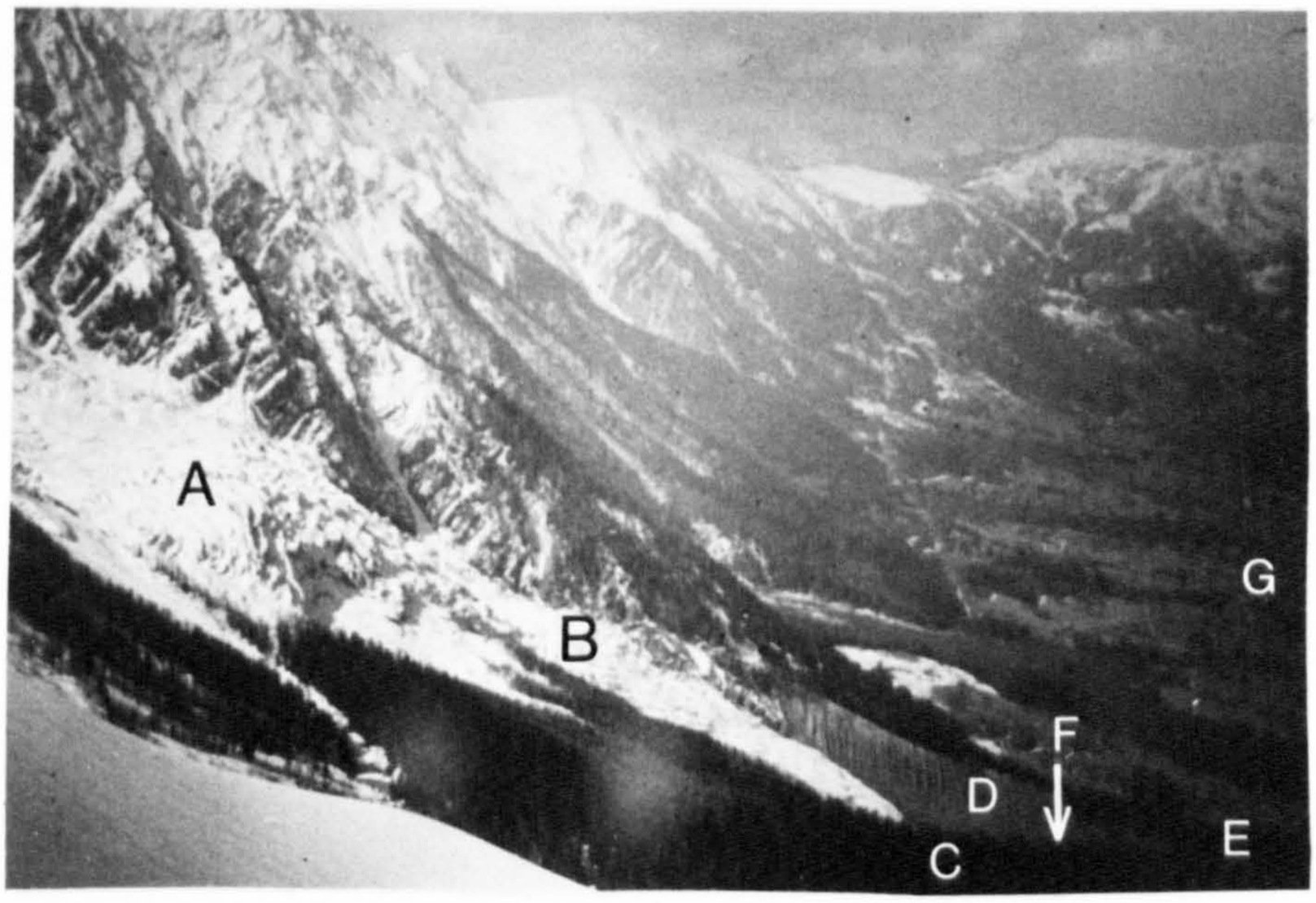
(2) Above 3000 m and at an equal distance from the ice-snout, the eastern névé lies up to 400 m higher than ice in the western névé. Below 3000 m, however, the eastern side of the glacier is consistently about 100 m lower than the western side. Ablation appears to be more efficient on the eastern side, unless the greater altitude in the eastern névé simply reflects a shallower cirque floor.

(3) The steep slopes of the massif are reflected in the overall steep gradient of the Bossons glacier, whose mean gradient is 24° (1 in 2.22, or 450m/km). It is one of the steepest glaciers in the French Alps, and as a result is also one of the fastest-moving. Its steepness, combined with its relatively short length, allows it to respond rapidly to variations in climate, and has proved a useful predictor of the variations in ice-snout movements of neighbouring glaciers. The glacier also extends to a lower altitude than any other glacier in the massif.

(4) The longitudinal profile tends to level out to form 'bench-like' features (Conard, 1931) at several altitudes (figure 2.4, page 58). According to Veyret (1959) these occur at 2800-3000 m, 2450-2620 m and at 1760-1840 m (see also Hougin, 1933). The alternating steep and level slopes of the glacier's longitudinal profile may largely reflect alternating, extending and compressing ice flow (e.g. Nye, 1952, Nye and Martin, 1968) which may have accentuated minor bedrock irregularities, weaknesses and constrictions. Other workers have postulated that the benches were caused by ice erosion at the granite-schist contact, by subaerial valley-side dissection during several proglacial cycles of erosion (e.g. Godefroy, 1948), or by increased erosion beneath successive junctions of the Bossons glacier with the former Arve glacier, during the Pleistocene (e.g. Veyret, 1958, 1959). Any benches are likely to be composite in origin. However, the benches have not been located or mapped with any precision, no accurate altitudinal

FIGURE 2.4

THE BOSSONS GLACIER VIEWED FROM THE EAST



Legend

A Bossons Glacier, bench *II*, 2620 m to 2450 m
B Bossons Glacier, bench *I* , 1840 m to 1760 m
C 'Eastern' lateral moraine
D 'Western' lateral moraine
E Torrent des Bossons
F Proglacial valley
G Chamonix valley

values have been obtained in the field, and the structural and compositional nature of each bench has not been investigated.

The longitudinal profile discussed so far has been that of the ice-surface. Data for the construction of the bedrock profile are unavailable, for thicknesses of ice have not been investigated. Veyret (1959) claims that the glacier is quite thin, and only flows over the mountain slope, although he provides no values of thicknesses. However, the form of the bedrock surface beneath the glacier cannot be judged solely from surface ice configuration and observed marginal ice thicknesses. Several other glaciers flowing into the Arve valley are underlain by deeply creviced bedrock and subglacially dissected gorges. Subglacial and seismic investigations have proved the existence of gorges beneath the Mer de Glace, the glacier d'Argentière and the Saint-Sorlin glacier. (Susstrunk, 1951; Lliboutry, 1962; Belin, 1962; Gluck, 1967.) A photograph of the Bossons glacier ice-snout taken during an ice retreat in 1952 shows areas of bedrock which are penetrated by an almost vertical, gorge-like trench. It therefore seems likely that the bedrock surface is relatively irregular, including gorge-like features, and cannot be determined directly from measurements of the ice-surface and of the regional topography.

2.3. CLIMATE

2.3.1. THE MONT BLANC MASSIF

The massif is prone to three climatic influences:

(a) 'Atlantic' air masses, penetrating from the west and northwest, bringing the warm, moist conditions characteristic of mid-latitude depressions and which are also the source of some winter snow.

(b) 'Continental' air masses, derived from northern and eastern regions, bring cold, strong winds such as the 'Bise'.

(c) 'Mediterranean', or southern, winds which are unstable and stormy. During the winter months the southerly winds bring 75 - 90 per cent of total snow amounts recorded at Chamonix.

Superimposed on the effects of the air masses are local conditions prevalent in the alpine mountain environment. These include variations of temperature and precipitation according to altitude. In the Mont Blanc massif, the negative temperature gradient ranges from 0.3°C to 0.5°C/100m in winter to 0.6°C to 0.7°C/100m in summer (Godefroy, 1948). The positive precipitation gradient is approximately 44mm/100m (Corbel, 1968), attaining a maximum at an altitude of 2700 to 3000 m (Blanchard, 1947), close to the permanent snowline. Aspect, exposure, local topography and the nature of any vegetative or ice cover all contribute to the local variability in microclimate, further modified by the development of valley and mountain winds.

Mean monthly temperatures at Chamonix[1] fall below freezing for three months of the year (December to February), and minimum temperatures fall to below -10°C on more than 30 days per annum. Snowfall during these three winter months accounts for only 19 per cent of total precipitation. Both maximum mean monthly temperatures (over 15°C) and maximum precipitation (33 per cent of total) occur during the three summer months of June, July and August. Total precipitation ranges between 1120 and 1269 mm p.a., with much summer rain derived from convectional currents set up over the Chamonix valley when ground-level air temperatures exceed 20 - 25°C.

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1. Climate data are available from several meteorological stations within the Chamonix valley, i.e. Chamonix La Tour (1460 m), Chamonix "Le Bouchet" (1037 m), and the Laboratoire d'Etude de la Recherche Scientifique, situated near the summit of the Aiguille du Midi (3613 m). Data have been collected since 1887, and some of them have been published by the "Département des Eaux et Forêts" (1912, 1925, 1933) and by the Météorologie Nationale (Paris)(1970). More detailed and more recent weather recordings have been made available by the P.S.H.M. (the mountain rescue police), Chamonix,for comparison with field data.

2.3.2. THE PROGLACIAL VALLEY OF THE BOSSONS GLACIER

Weather observations were made in the proglacial valley of the Bossons glacier in April and July 1971. The details of measurements and nature of the microclimate have been published elsewhere (Maizels, 1973, and in preparation, a).

2.3.2.1. Proglacial temperatures

Hourly temperatures and temperature range were recorded on nine maximum-and-minimum thermometers and a thermohygrograph placed at selected distances from the ice-margin on the valley-train, and at the foot of each lateral moraine. The thermometer sites are plotted on the base-map surveyed by the author (figure 2.5, page 62). During the two periods, over 85 temperature recordings were obtained from each instrument. Thermometers were fixed at 20 cm above ground-level, on a one-metre pole, facing due north. Station 3 included (a) a Stevenson screen containing thermohygrograph and wet-and-dry bulb thermometers, (b) a maximum-and-minimum thermometer as at the other 'stations', and (c) a rain gauge (figure 2.6, page 63). Errors in measurements of temperature, based on comparison of thermometer recordings with thermograph measurements at station 3, are less than 0.5°C for temperatures between 1°C and 18°C, up to 2°C for temperatures greater than 18°C, and up to 3°C for temperatures less than 1°C (to -10°C).

A distinct rise in temperature was recorded with increasing distance from the glacier (table 2.1, page and figure 2.8, page 65). The proglacial temperature rise is steepest within about 40 m of the ice-margin, after which temperatures tend to level out to 200 m from the ice-margin. Lower temperatures were recorded from the distal end of the valley-train where glacier winds become channeled through the valley exit. Day-time valley winds, blowing up from the Chamonix valley, periodically modify the proglacial wind system operating over the distal part of the valley-train (see Geiger, 1965).

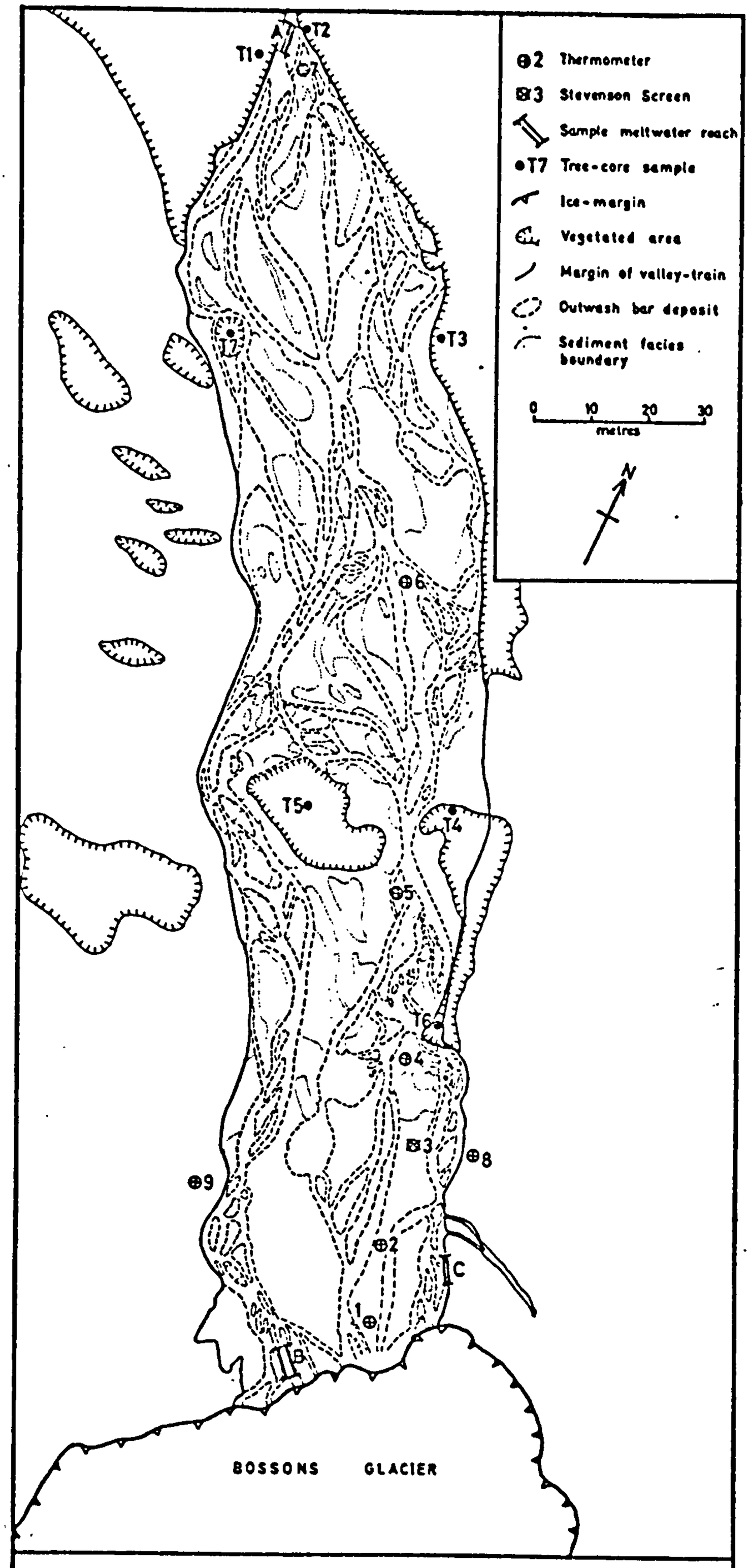


Figure 2.5 Meteorological, hydrological and dendrochronological sites on the Bossons valley-train

(Surveyed in April, 1971)

Figure 2.6
Thermohygrograph site on the Bossons proglacial valley-train, July,1971

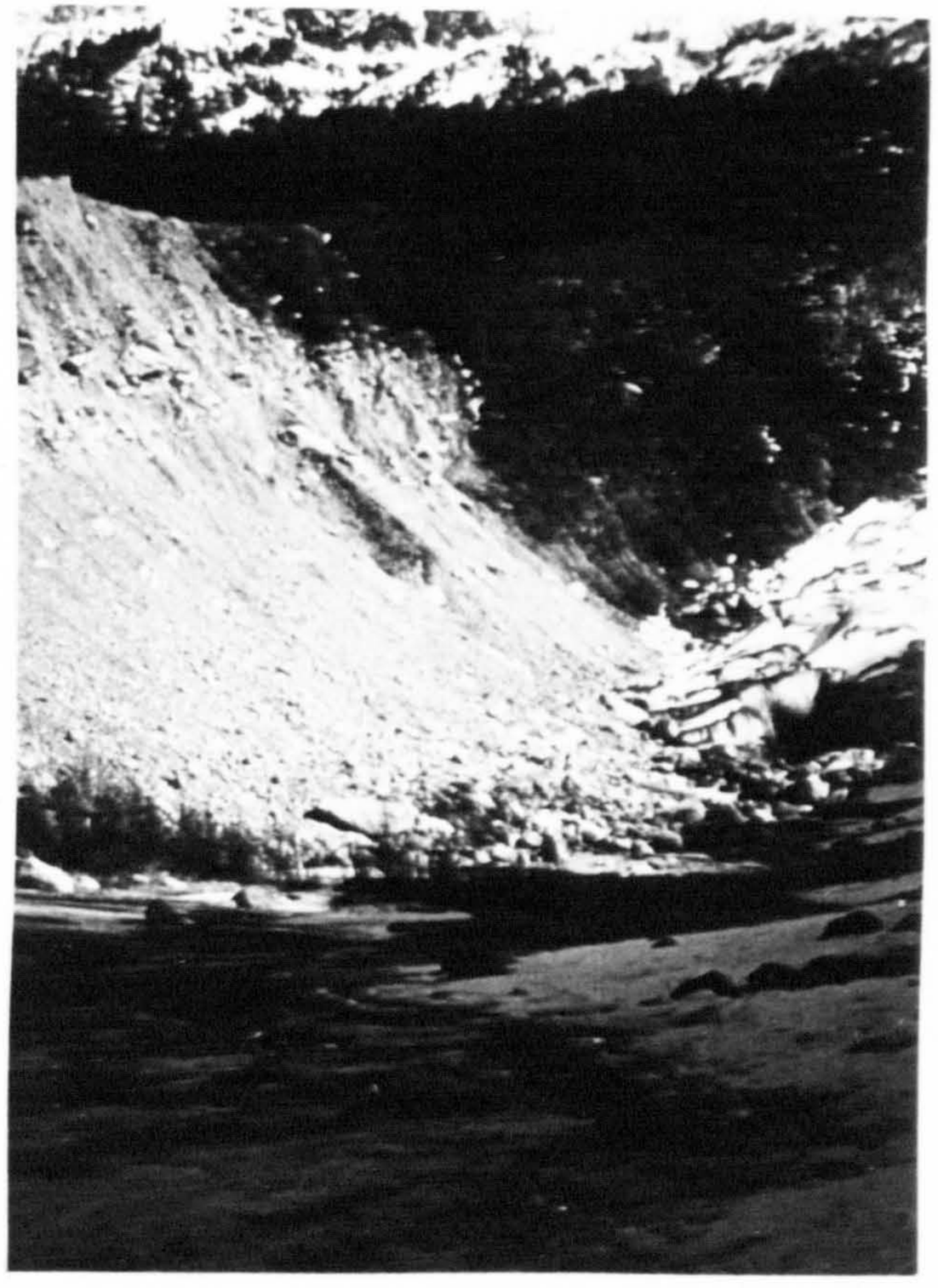


Figure 2.7
Differential insolation in the Bossons proglacial valley

Table 2.1

Mean temperature and temperature range in the Bossons proglacial valley and at Chamonix, April and July, 1971

Thermometer site (see fig. 2.5)	April				July			
	Distance from ice-margin (metres)	Mean temperature (°C)	Mean maximum temp. (°C)	Mean minimum temp. (°C)	Distance from ice-margin (metres)	Mean temperature (°C)	Mean maximum temp. (°C)	Mean minimum temp. (°C)
valley-train:								
1	6.0	6.64	9.61	2.36	6.6	10.47	18.24	6.69
2	18.0	7.29	10.58	2.00	18.0	11.24	15.64	7.35
3	33.0	7.12	9.68	2.61	40.8	12.20	16.83	7.63
3T	33.0	7.19	10.53	3.84	40.8	12.67	17.74	8.93
4	48.0	7.46	11.28	1.78	43.8	12.08	16.19	7.39
5	76.5	7.74	11.25	1.64	75.6	12.02	16.56	8.51
6	127.5	8.01	10.50	2.39	111.6	13.33	18.69	8.04
7	217.5	4.28	8.61	-2.03	218.7	8.58	15.23	3.26
lateral moraines:								
8	30.0	7.24	10.69	1.26	14.4	13.26	18.37	8.21
9	40.5	7.27	10.17	1.83	40.2	13.41	19.22	7.91
Chamonix		9.50	13.44	3.67		16.50	23.63	9.63

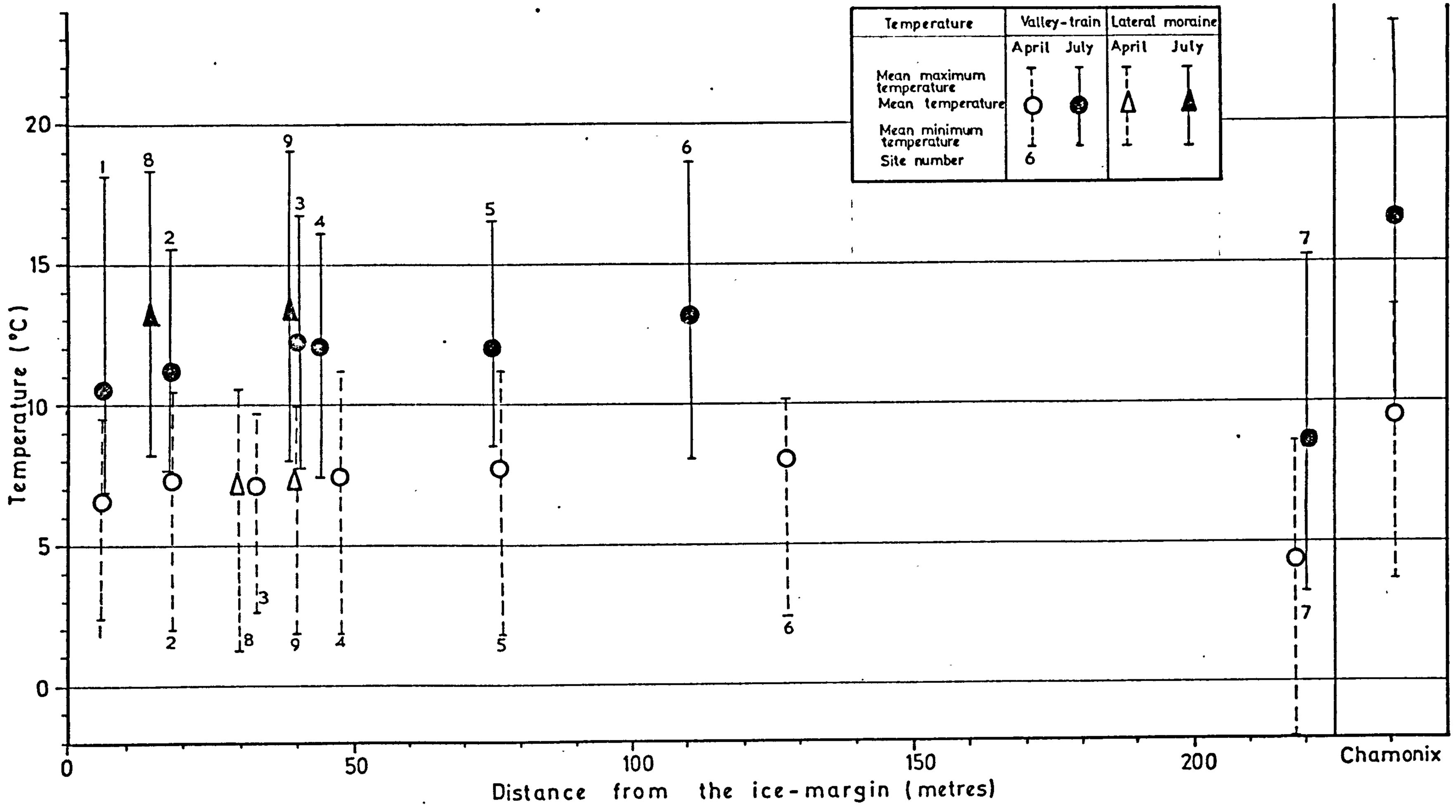
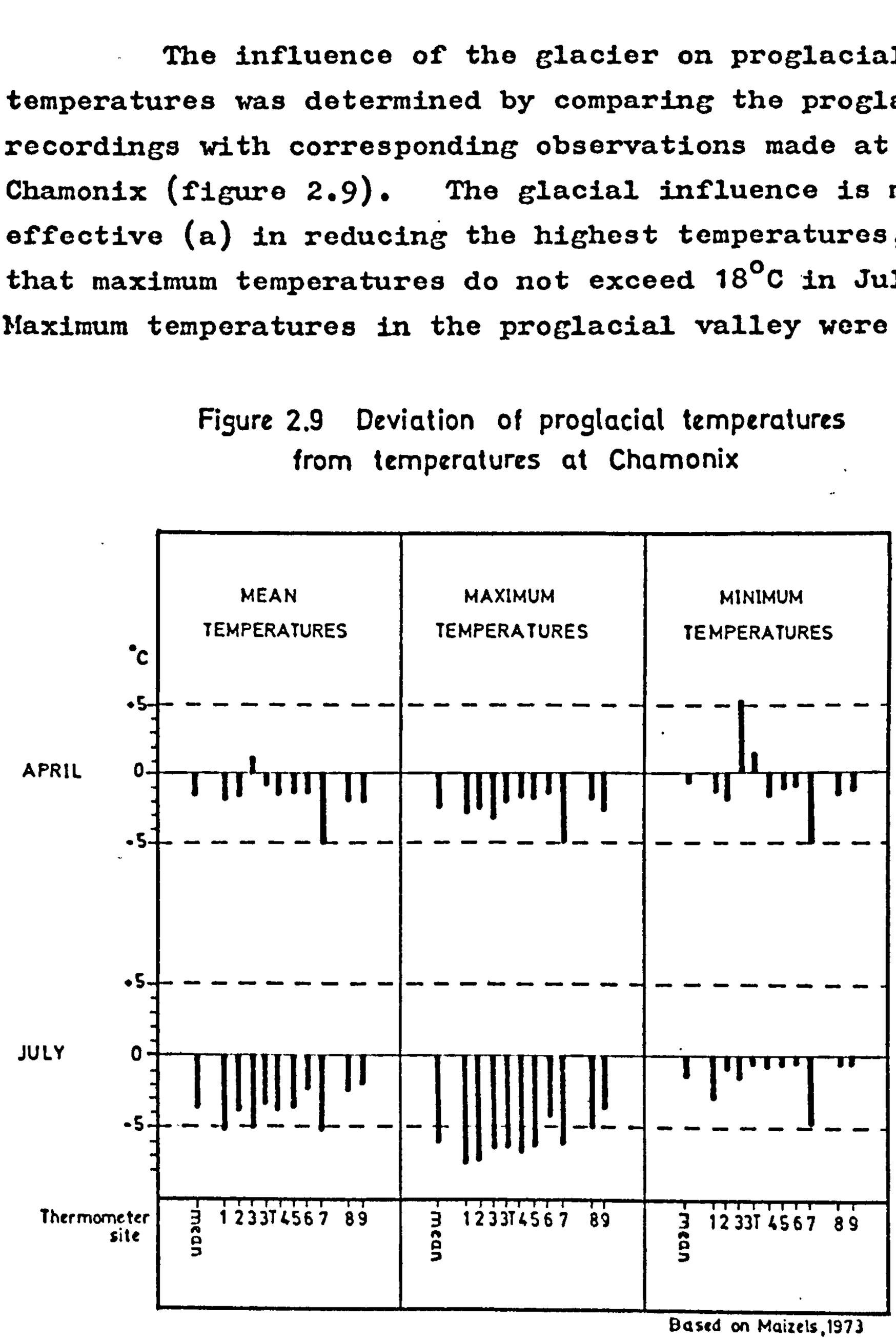


Figure 2.8 Temperatures in the Bossons proglacial valley and at Chamonix, April and July, 1971

Daily weather variations affect proglacial temperatures, far greater variations in diurnal and downvalley temperatures occurred during periods of clear weather (see also Lindroth, 1963 and Benford, 1970). The downvalley temperature gradient reached a maximum of 2.7°C/100m in July 1971 when the maximum recorded temperature in the valley was about 18°C. Zero temperature gradients were recorded during cloudy or wet weather conditions when maximum temperatures lay between 2°C and 8°C; a negative downvalley temperature gradient developed during colder periods when the temperature of the glacier wind was higher than that of the main valley winds.

The influence of the glacier on proglacial temperatures was determined by comparing the proglacial recordings with corresponding observations made at Chamonix (figure 2.9). The glacial influence is most effective (a) in reducing the highest temperatures, so that maximum temperatures do not exceed 18°C in July. Maximum temperatures in the proglacial valley were reduced,

Figure 2.9 Deviation of proglacial temperatures from temperatures at Chamonix



Based on Maizels, 1973

on average, by 2.3°C in April and 6.05°C in July, compared with the maximum 'expected' temperature reduction resulting from differences in altitude of 1.25°C; and (b) in preventing minimum temperatures from falling much below 0°C. Minimum temperatures in the proglacial valley often deviate from those at Chamonix by less than 1°C, and may even be higher than at Chamonix, especially during the winter. The proglacial valley experiences a smaller annual and diurnal temperature range than does Chamonix as a direct result of cold, often strong, winds blowing off the Bossons glacier towards the Chamonix valley. The winds are strongest in summer, when pressure and temperature differences between the proglacial valley and the main valley are greatest (see also Wojcik, 1973a). Proglacial air temperatures are, therefore, further reduced in summer as glacier winds bring in a constant supply of cold air to the valley.

2.3.2.2. Humidity and rainfall

Relative humidity is significantly lower in the Bossons valley than at Chamonix, corresponding partly to the lower temperatures occurring in the proglacial valley, and partly to the abundance of moisture in the nearby ice-mass and the numerous meltwater streams.

The local climate is also affected by the aspect and degree of shelter of the proglacial valley. The valley is aligned north-northwest, overlooked by steep lateral moraines. The 'eastern' lateral moraine receives more direct solar radiation than the opposite slope, the 'western' lateral moraine, particularly in the late after-noon, and also experiences greater extremes of temperature. As a result, snowmelt is more rapid and spring snow-lie is shorter on the 'eastern' lateral moraine than on the oppposite slope (figure 2.7, page 63). The number of frost cycles, especially during spring and autumn (e.g. Troll, 1943) is probably more frequent on this slope.

The valley is also relatively sheltered from the prevailing south and southwesterly winds and the associated precipitation, by the surrounding high mountain ridges. Rainfall over a nine-day period in July, for example, totalled 30.00 mm at Chamonix, but only 8.55 mm in the Bossons proglacial valley. If the "precipitation-gradient with altitude" of 44 mm/100 m (Corbel, 1968) were applicable, the Bossons valley should have received over 100 mm during that period. The Bossons valley therefore appears to be unusually sheltered from regional air masses, and its climate is, instead, dominated by local winds blowing from the glacier.

2.3.2.3. Microclimatic zones

Three microclimatic zones may be recognized in front of the ice-margin, and a fourth, more distant one, may be inferred.

Zone I lies 0-40 m from the ice-margin; the influence of cold air blowing directly off the ice is at a maximum, although it sharply declines with increasing distance downvalley. Lindroth's (1963) zone of maximum influence extended to about 30 m.

Zone II lies between 40 and 200 m from the ice-margin; the influence of cold glacial air is substantial, but varies with changes in seasonal weather conditions and the degree of penetration by valley or regional winds.

Zone III lies further than 200 m from the ice-margin; the 'glacial' influence is only important intermittently, i.e. when glacial winds are strong. The relative extents of zones I and II appear to increase exponentially. Assuming this pattern is also extended to zone III, this latter zone extends to about 1 km from the ice-margin.

Zone IV lies further than 1 km from the ice-margin. Here the climate is rarely influenced by ice-cooled air masses, and is instead governed largely by the 'non-glacial' mountain climate of the massif. Zone IV would extend about 5 km from the ice-margin.

The maximum distance at which the Bossons glacier regularly affects its proglacial climate is about 1 km, but the proglacial climate is nevertheless very closely related to the regional climate and only slightly modifies it. The proximity of other glaciers in the region does not appear to affect the local climate for the glaciers are relatively small, and their effect on altering the regional climate is slight, whereas the influence of the regional climate is great. Thus the larger the glacier, the greater the influence on its proglacial climate, and the less important is the regional climate. Once a glacier attains a particular size, the effect on proglacial climate is so profound and extensive that the former "non-glacial", regional climate is replaced by a "glacial", regional climate. The severity of the proglacial climate therefore depends on the interaction between

(1) the size of the glacier or ice-mass, and
(2) the initial regional climate (particularly temperature conditions).

2.4. GLACIATION AND STUDIES OF GLACIER VARIATIONS

2.4.1. PREVIOUS GLACIOLOGICAL STUDIES IN THE MONT BLANC MASSIF

2.4.1.1. Preliminary observations

The first scientific observations on the Mont Blanc glaciers were undertaken in the late eighteenth century. H.B. de Saussure (1760-1781) for example, was one of the first to associate moraines with former glacier advances. Some of the most important achievements in glaciology and glacial geomorphology were accomplished during the nineteenth century. The now-popular resort of Chamonix, overlooked by five important and relatively accessible glaciers, provided a starting-point for much pioneering glaciological research. Venetz (1833, 1861), Agassiz (1840, 1847) and Forbes (1843, 1853, 1857, 1865) described in detail many glacial features of the massif and carried out a great number of measurements on glaciers to help determine the

thickness, temperature, rates of ablation and movement of the ice. By the 1880's an enormous amount of information had been gathered about the glaciers of the massif, which provided the bases of some widely accepted concepts in glaciology (summarized for example, by Viollet-le-Duc, 1877) and in glacial geomorphology (e.g. 'the cycle of glacial denudation' of Davis (1900)).

2.4.1.2. Measurements of ice-snout variations

By the 1890's interest had turned to determining the likely causes of changes in glacier volume and the position of the ice-snout. Investigations therefore concentrated on seasonal measurements of the ice-snout position, with the aim of relating any variations to regular meteorological observations.

In 1891, Bonaparte drew attention to the rapidity with which the position of an ice-snout may change, and he took regular systematic measurements of ice-snout positions with reference to marked boulders, and determined some historical variations from old documents, maps, drawings and photographs published since the sixteenth century (e.g. figure 2.10, page 71). Rabot (1920a,b) drew up a comprehensive summary of Alpine glacier variations from the end of the sixteenth century, and compared these with glacier variations recorded in Alaska, Iceland, Norway and Spitsbergen. His results were also based on frequent and regular observations by Payot, J. and C. Vallot, Forel (working mainly in Switzerland) and Fantoli (in Italy), who had determined the ice-snout positions by the use of both trigonometry (theodolite and plane-table) and photogrammetry (using 5000 'geometric' photographs). In 1911, the Vallots produced a map to show the extents of at least nine Mont Blanc glaciers, at a scale of 1:20 000. They also measured the lengths and gradients of the glaciers, estimated the rate of ablation of the Mer de Glace over a hundred year period, and summarized all the information known on annual movements of the Mont Blanc glaciers (1909, 1922). More precise measurements have been obtained more recently, either by direct surveys (e.g. Mougin, 1909, 1912, 1925, 1933;

Figure 2.10 The Chamonix valley in 1787 A.D. (after Hackert)
1 Bossons glacier 2 Mer de Glace

Mercanton, 1948, 1950, 1952, 1953, 1957) or by photogrammetric methods (e.g. Finsterwalder, 1959; Baussart, 1960; I.A.S.H., Commission of Snow and Ice).

However, variations in glacier length and the position of the ice-snout had gradually come to be regarded as less accurate indicators of variations in glacier mass than were changes in the level of the whole ice-surface. While some workers still reported on recent and historical glacier variations (e.g. Blanchard, 1913; Allix, 1922) and on glacier-related phenomena such as floods (e.g. Jourdan-la-Forte, 1920), many other workers turned back to investigating the internal physical conditions which produced changes in ice volume of the glaciers, as initiated by the glaciological observations of Agassiz.

2.4.1.3. Investigations of glacier mass budgets

As early as 1905, Mercanton (1905) had described some possible methods of drilling into a glacier to measure its internal temperature and velocity, in order to throw light on how the glacier actually moved, and at what rate meltwater was produced or filtered through the glacier. One of the chief incentives promoting investigations into the internal physics and structure of Mont Blanc glaciers, once Mercanton had proposed the methods, was the demand for hydro-electric power, supported by l'Electricité de France (see Sylvestre (1922), and Waeber (1943)). A series of gauging stations was established and meltwater discharges were regularly measured on the major outlet streams from the Mont Blanc glaciers.

By the late 1940's, glaciological and hydrological investigations had succeeded in measuring internal temperatures and rates of ice movement; subglacial investigations in the Mont Blanc massif had become well advanced, and huge quantities of glacial meltwater harnessed. In addition, historical glacier variations had become well-documented, and described in detail by Kinzl (1932), Von Klebelsberg (1943), Billwiller (1948), Onde (1948), Mercanton (1948, 1951, 1957), Mercanton and Renaud (1949, 1950, 1952, 1953) and Sourbier (1950).

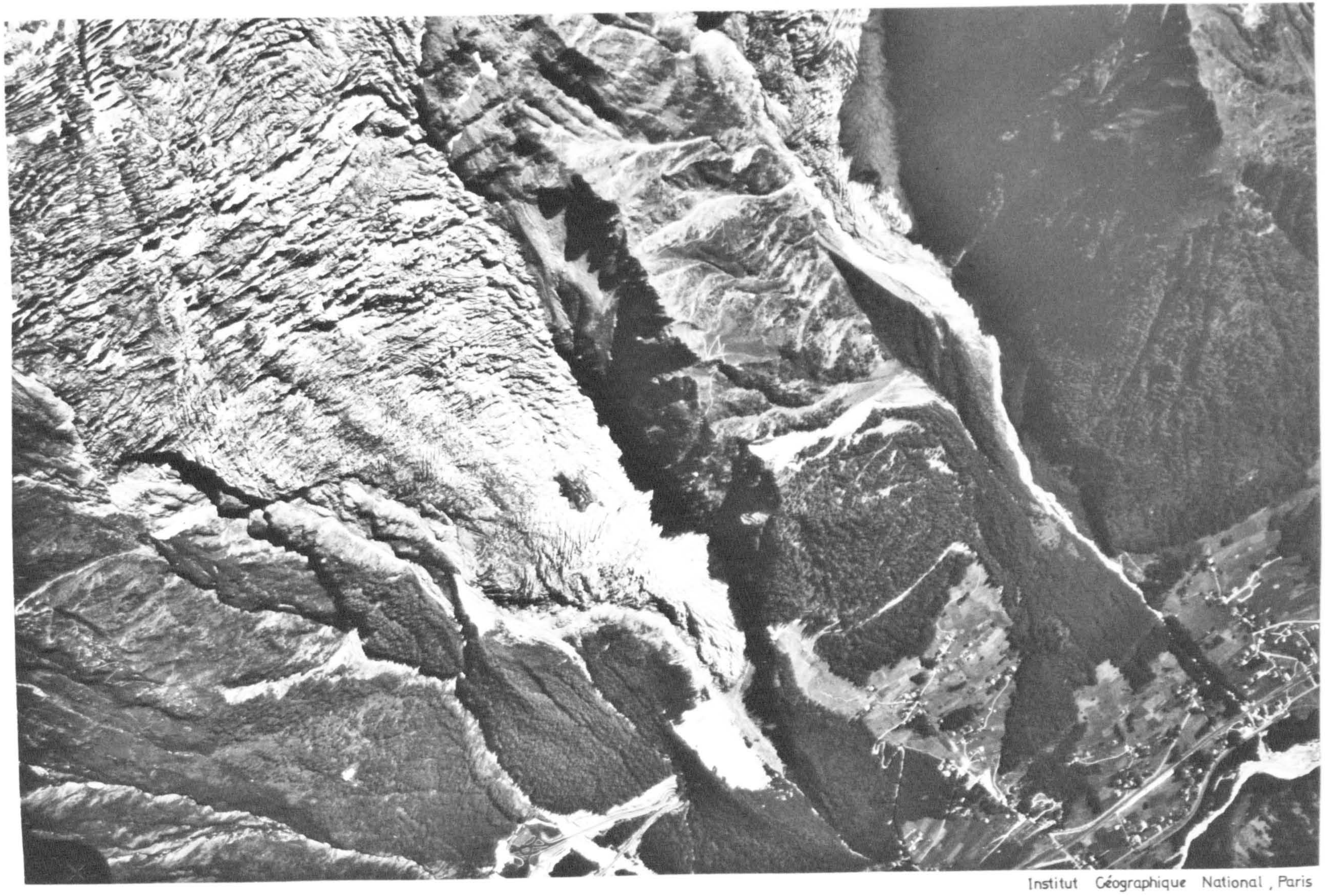
2.4.1.4. Applications of new techniques

Some important new techniques were introduced to glaciological studies in the Mont Blanc massif after the second World War. The new techniques included seismic surveys, aerial photography, numerical techniques for predicting glacier variations from climatic data, and radiocarbon dating.

Renaud and Mercanton (1948) were the first to use a seismic survey to determine ice thickness over a glacier basin, although originally suggested by Mercanton in 1926. Susstrunk (1951), Ract-Madoux and Reynaud (1951), Lliboutry and Vivet (1961), Belin (1962) and Gluck (1967) all carried out seismic surveys to determine ice-thicknesses and the shape of the subglacial bed. The Mer de Glace attracted most workers, and was found to be floored by a deep trough which was 'V-shaped' according to Susstrunk (1951) and 'U-shaped' according to Lliboutry (1962), while other bedrock floors were found to be deeply creviced (Belin, 1962).

Aerial photography has proved very important in the analysis of rates of glacier movement and changes in surface height and ice-marginal positions of the Mont Blanc glaciers (e.g. figure 2.11, page 74). It has also been used as the basis of recent topographic maps published by the Institut de Géographie Nationale (e.g. Feuille Argentiere, 1:10000, 1959).

Millecamps (1957, and see 1959) measured rates of ice surface lowering between the 1939 and the 1952 French airforce series of aerial photographs. Several photographic series were also used by Baussart (1960; Baussart et al., 1965) to deduce the changes in ice volume in the accumulation and ablation zones of several of the Mont Blanc glaciers. Measurements of surface ice velocity were based on the downvalley movement of ogives, which are easily recognized on these aerial photographs. Lliboutry (Lliboutry et al., 1962) combined the interpretation of the aerial photographs with tacheometric and seismic surveys, to produce a map which he could compare with Vallot's 1911 map. He then estimated changes in annual accumulation and ablation over the period 1911 to 1958.



Institut Géographique National, Paris

Figure 2.11 Aerial photograph of the Bossons glacier snout and its proglacial valley, 1966

The relationship between climate and glacier variations has been tackled quite recently by Vivian (1973) who used a multiple regression equation to predict past ice-snout variations of the Mer de Glace. Prediction was based on the position of the ice-snout in the previous year, the precipitation of the preceding winter, and mean summer temperature (June to August). But variations in the position of the ice-snout are less reliable measures of the glacier's response to climatic conditions than are mass balance or volume measurements based on amounts of ablation and accumulation over the whole glacier (Paterson, 1969). Each glacier will respond differently to climatic conditions, according to the site, altitude, length, slope and exposure. The relative importance of each of Vivian's factors therefore remains undetermined in relation to other glaciers.

Peguy (1947) derived a descriptive equation relating the altitude of the ice-snout to the local snow-line, the height of the massif, the area, orientation and gradient of the glacier. It may be useful to build some of these physical factors into Vivian's equations in order to extend predictions to neighbouring and other glaciers, particularly if mass balance data are lacking.

Corbel and Leroy Ladurie (1963) were the first to employ radiocarbon dating techniques on any Mont Blanc deposits, although used earlier at the Aletsch Glacier (Oeschger and Rothlisberger, 1961). Pieces of wood from a major lateral moraine overlooking the Taconna Glacier (see figure 2.11, page 74) were dated to 1630 and 1680 AD. (approximately). The moraine was probably formed during a glacial advance recorded in the mid-seventeenth century (1643-44 AD.) and modified by a further advance in 1723 AD. Mayr (1969) and Lamarche and Fritts (1971) also used biological evidence as evidence of the dates and extents of postglacial glacier variations. In addition to analyses of humus layers and buried soils on and in moraines, Mayr investigated varves formed in an ice-marginal lake at the Miage Glacier, Val Veni (Italy). He attributed probable dates to the various horizons sampled, which corresponded

with dates originally proposed elsewhere in the literature.

Descriptions of historical glacier variations have also persisted into the more recent literature. Vanni (1957, 1962, 1967) and Corbel (1963, 1968) related glacier variations to climatic data and attempted to recognize cycles of glacial advances. Bouverot (1957) and Finsterwalder (1959) have described the historical glacier variations of individual glaciers of the Mont Blanc massif, and suggested possible causes of advances. Fairbridge (1961) and Grove (1966) presented general reviews of glacier variations based on historical evidence. Many of the most recent investigations have been summarized by Vivian (1971).

2.4.2. PLEISTOCENE GLACIATIONS

Penck and Brückner (1909) developed the fundamental framework for the Pleistocene glacial chronology. This comprised at least four glaciations, named the Gunz, Mindel, Riss and Würm Glaciations, probably preceded by the Donau Glaciation (e.g. Segota, 1966). The divisions were based on the morphology, weathering and stratigraphical position of a series of outwash terraces, the Deckenschotter and Terrassenschotter near Memmingen in the eastern Alps (Zeuner, 1935, 1959; Godefroy, 1948; Flint, 1957; Richmond, 1972). The Würm comprised three major phases, designated Würm I, Würm IIa and Würm IIb (Troll, 1926; Eberl, 1928; Knauer, 1928; Beck, 1937, in Zeuner, 1959; but see Brunnacker, 1975). Workers in different parts of the Alps have provided different series of names for each set of deposits, and correlation both within the Alps and with extra-Alpine areas has not yet been completed. In addition, the extent of each Würm glaciation in many areas has not been agreed upon.

The highest peaks of the Mont Blanc massif stood above the ice as nunataks during the maximum glacial phases of the Pleistocene (Penck and Bruckner, 1909; Flint, 1957; Corbel, 1968). Evidence of nunataks is based on the upper limit of glacially-smoothed valley-sides ("la limite des

polis glaciaires" of Godefroy, 1948). According to Corbel (1968), the surface of the Arve Glacier lay at 2300 m during Würm I (15000 to 13000 years B.P.) and at 1450-1500 m during Würm II (8000 years B.P.). This suggests that the nunatak peaks projected between 2500 and 3350 metres above the ice during the last glacial phases, and that ice thicknesses exceeded 1300 metres.

Little additional evidence remains in the Chamonix valley of Pleistocene glaciations, for deep erosion and subsequent valley-floor infill between confined valley walls has obliterated any glacial deposits.

2.4.3. POST-GLACIAL AND RECENT GLACIER VARIATIONS

2.4.3.1. The Mont Blanc massif

Four postglacial cold phases have been identified in the Alps: The Egesen (or Daun) of the Late Glacial; the Larstig phase which occurred during pollen zone VI (the Late Boreal), possibly extending into the hypsithermal climatic optimum (about 8500 years B.P.) or pollen zone VII (the Atlantic); the Simming, which occurred early in pollen zone IX (the sub-Atlantic); and finally, the Fernau, or "Little Ice Age" ("Frührezent" of Kinzl, 1932), which lasted from 1600 AD. to 1850 AD. (Mayr, 1969; Grove, 1966). Mayr (1969) has identified and mapped the former frontal moraines of the major glaciers in the massif. Material from buried soils in the oldest moraine was radiocarbon dated to 8400 years B.P. $\pm$100 years, i.e. the Larstig cold phase. Moraines dating from the Simming phase have been identified in front of most glaciers, but the best-preserved glacial deposits are those dating from the most recent advances. Postglacial glacier variations since 1600 A.D. have attracted much attention in the literature (section 2.4.1.). A detailed and precise chronology has been worked out for the "Little Ice Age" period, and is summarized in table 2.2. (page 78).

Table 2.2

Historical glacier variations in the Mont Blanc Massif

[see text for sources of information]

SIXTEENTH CENTURY	
1580	Records of glacier advances with violent avalanches.
1595-1601	Records of advances and floods from four glaciers in Switzerland and Austria to produce moraines.
SEVENTEENTH CENTURY	
1605	Chamonix population concerned about loss of land resulting from glacier advances and meltwater floods. Daily advance of Argentière glacier reported.
1610-1616	Advance of Mer de Glace; meltwaters change course on floor of Chamonix valley.
1616-1630	Retreat or stationary glaciers.
1630-1640	Small advances, and 5 floods recorded.
1641-1643	Advances reported, menacing of villages by meltwaters, and destruction of cultivated lands. Formation of a large moraine.
1643-1670	Glaciers retreat.
1663	Land exorcised to prevent overriding of land by ice.
1677	Glacier advances.
1679-1680	Floods reported.
1680	Advance of glaciers; formation of moraines?
1690	Glaciers retreat.
1694	Some glaciers reported to be advancing.
EIGHTEENTH CENTURY	
1703	Some glaciers reported to be advancing.
1716-1724	Some glaciers reported to be advancing.
1723	Glacier advances, formation of large moraines.
1728-1732	Cadastral survey shows extended glacier snouts.
1740-1743	Small glacier advances.
1770-1780	Large glacier advances.

continued..

Table 2.2. continued

NINETEENTH CENTURY	
1818-1820	Major glacier advances and formation of large moraines.
1825	Mer de Glace still in advanced position.
1830	Glacier retreat begins.
1850	Glacier advances, formation of moraines.
1875	Small glacier advances.
1883-1896	Glacier advances recorded.
1895-1915	Slow glacier retreats.
TWENTIETH CENTURY	
1910-1925	Glaciers advancing.
1925-1938	Glaciers retreat.
1938-1941	Glaciers advancing.
1941-1953	Glaciers retreat.
1953-1974	Glaciers stationary or advancing.

Five 'major' glacial advances took place in the Mont Blanc massif between 1580 and 1850 A.D. These occurred in 1580-1610, 1640, 1680-1730, 1818-1820 and 1850. Several prominent terminal and lateral moraines are associated with these advances, the largest moraines dating from the mid-seventeenth century advance (Corbel and Leroy Ladurie, 1963), and smaller moraines from early nineteenth century advance (Kinzl, 1932). A small glacier advance in the 1890's was followed by a retreat, and a second advance culminating in 1920. The subsequent retreat was again followed by an advance in the early 1940's, which then rapidly retreated until 1952-53. However, since the early 1950's most of Mont Blanc glaciers exhibited no continuous trend until 1970, when 18 per cent of French glaciers were recorded to be advancing (Vivian, 1973).

The major glacier advances in the massif correlate fairly closely with advances recorded elsewhere in the world, including North America (e.g. Meier, 1965; Porter and Denton, 1967), Iceland (e.g. Eythorsson, 1949; John and Sugden, 1962; Thorarinsson, 1964), Greenland (e.g. Weidick, 1959), Norway and Sweden (e.g. Ahlmann, 1922; Faegri, 1948; Thorarinsson, 1940, 1964; King, 1959; Liestøl, 1960), the Himalayas (e.g. Thorarinsson, 1940) and New Zealand (e.g. Suggate, 1950), as well as other parts of the Alps, (e.g. Hoinkes, 1969; Röthlisberger, 1969).

2.4.3.2. Recent variations of the Bossons glacier

Numerous workers have discussed the ice-snout and moraines of the Bossons glacier, including Forbes (1843), Bonaparte (1891), Blanchard (1913), Kinzl (1932), Bouverot (1957) and Finsterwalder (1959). The large amount of data available on ice-snout variations of the Bossons glacier have been compiled here into a map (figure 2.12, page 81) and a graph (figure 2.13, page 82) showing variations since records began.

Distances are measured between the ice-snout and the profile R_1-R_1 bis, established by Mougin (1912). A second, higher profile was established in 1950 as a base-line for ice-margin measurements during glacier retreat (Bouverot, 1957). Measurements since 1968 are based on the author's own surveys (Table 2.3.). The Institut de Géographie Alpine at Grenoble have also been monitoring recent movements, but distances have been measured only in relation to reference boulders.

Advances have occurred in 1818, 1854, 1892, 1920, 1942 and post-1952. The glacier snout advanced nearly 500 metres between 1952 and 1974, although its rate of advance has declined since 1971. Most of these advances correspond with others recorded in the massif, and also with advances occurring in other glaciated regions of the world.

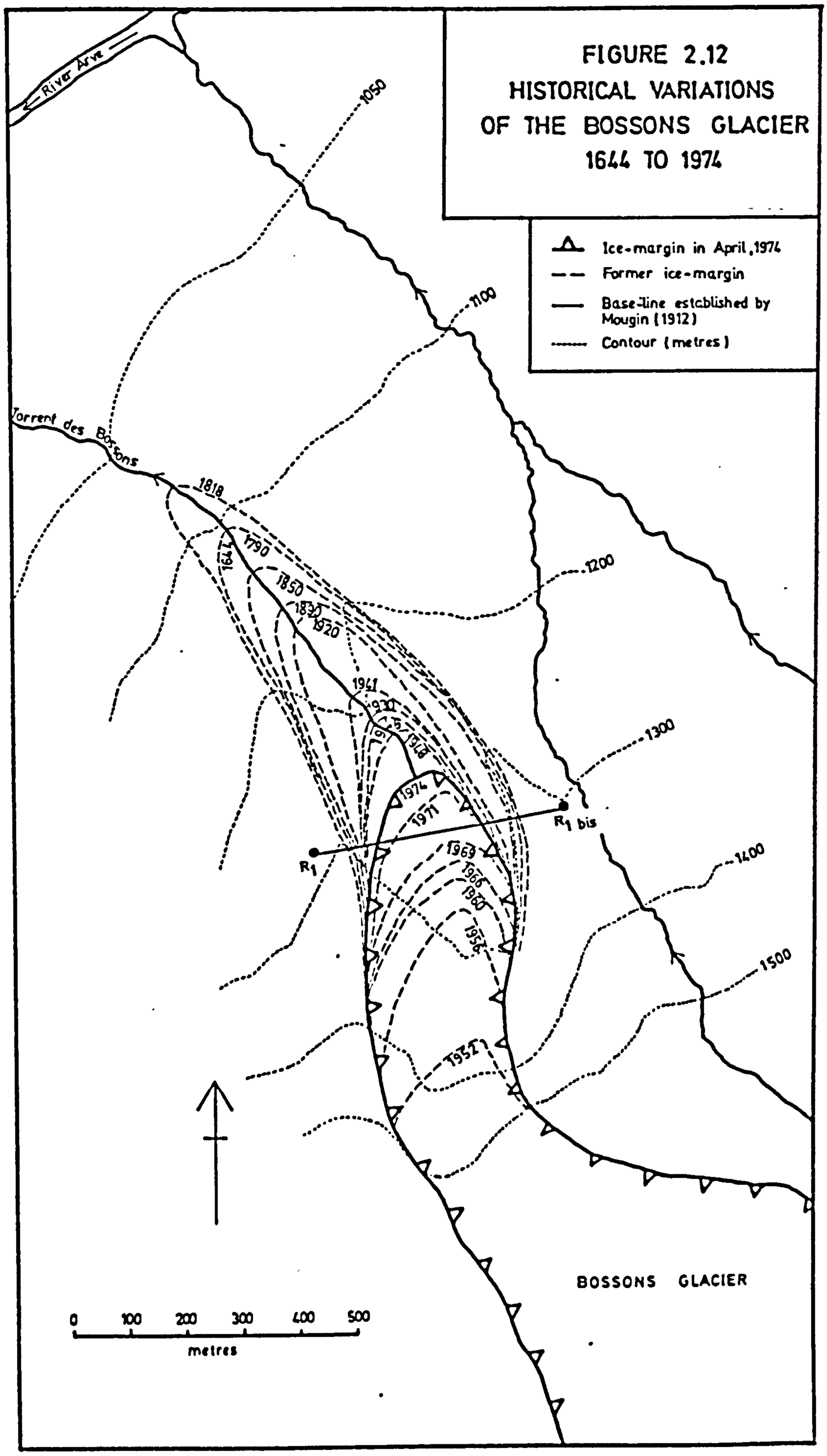


FIGURE 2.12
HISTORICAL VARIATIONS
OF THE BOSSONS GLACIER
1644 TO 1974
Ice-margin in April, 1974
Former ice-margin
Base-line established by Mougin (1912)
Contour (metres)
River Arve
Torrent des Bossons
1050
1100
1200
1300
1400
1500
1818
1790
1644
1850
1835
1920
1941
1930
1957
1913
1974
1971
1963
1966
1960
1954
1932
R1
R1 bis
0 100 200 300 400 500
metres
BOSSONS GLACIER

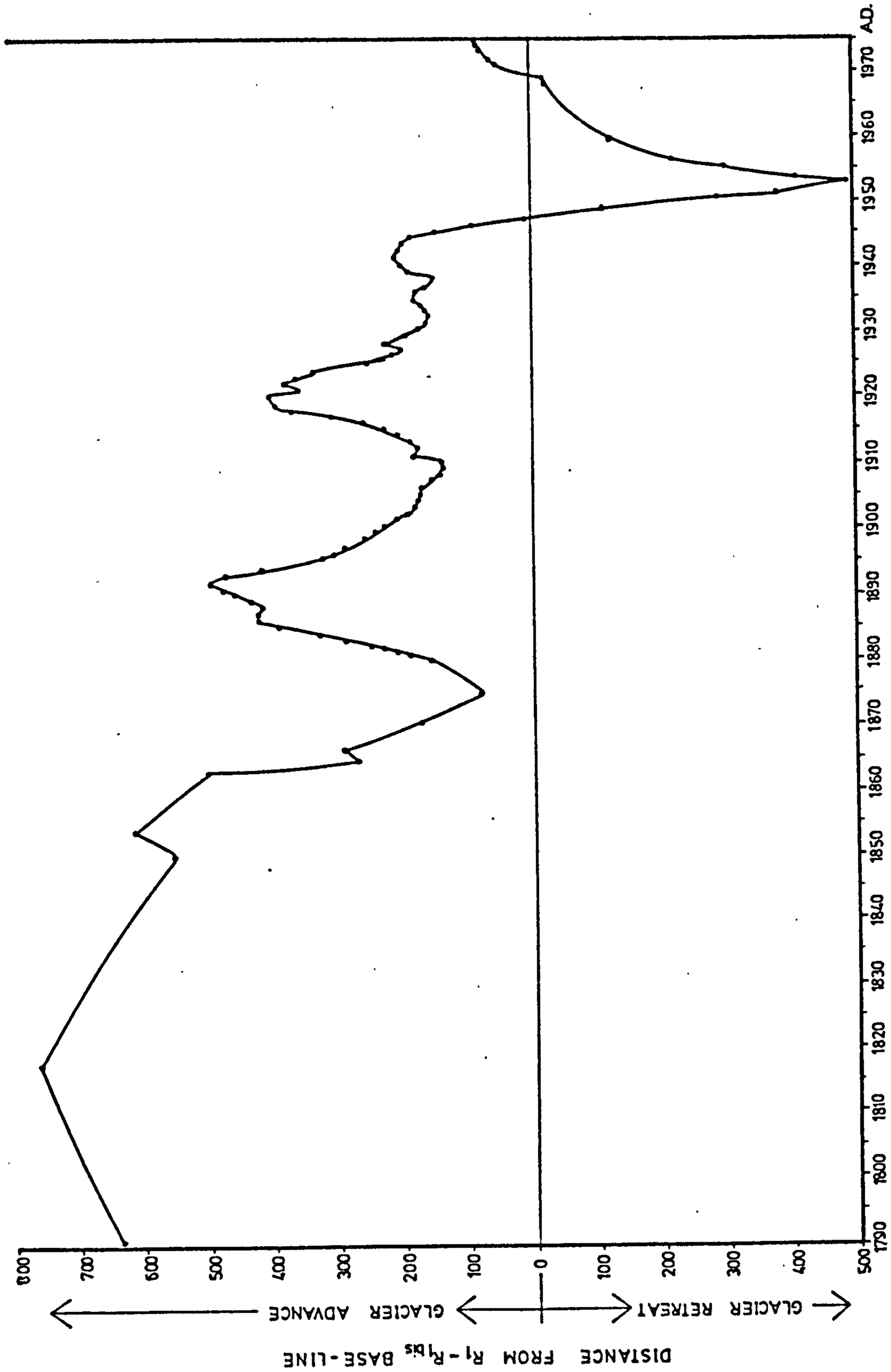


Figure 2.13 Fluctuations of the Bossons glacier snout, 1790 to 1975
(Based on Maizels 1970; see text for other sources)

Table 2.3

Position of ice-snout from R_1-R_{1bis} profile 1968-1974
(based on field surveys)

Year	Month	Distance (metres)	Mean monthly rate of change (metres)
1968	August	-24*	
1969	August	-21	+0.25
1971	April	+54	+3.57
1971	July	+61	+2.33
1973	April	+82	+0.95
1973	September**	+80	-0.50
1974	April	+86	+0.75
1974	September**	+82	-0.95
1975	June	+83	+0.11

** Personal communication

* Negative values indicate that the ice-snout lay upvalley of the profile line.

2.5. HYDROLOGY

2.5.1. THE MONT BLANC MASSIF

2.5.1.1. Measurements of meltwater discharge

The discharge of meltwater streams from five glaciers on the north-facing slope of the Mont Blanc massif has been measured by l'Electricité de France[1], using automatic limnographs. The data available include mean daily, monthly and annual discharge values at five gauging stations for the periods indicated in table 2.4 (page 84).

1. Data made available by M. Siegfried of E.D.F.-R.E.H. Alpes-Nord, Chambéry.

Table 2.4

Dates of discharge measurements available for Mont Blanc glaciers

Name of Glacier	Altitude of gauging station	Period for which measurements are available
Mer de Glace	1059 m	July 1950 - Dec. 1970
Argentière	1400 m	Jan.1955 - Sept.1970
Tré-la-Tête	1420 m	Aug. 1957 - Dec. 1968
Bionnassay	1070 m	May 1959 - Dec. 1968
Tour	1475 m	Aug. 1960 - Oct. 1967

2.5.1.2. Annual régime

The discharges of the meltwater streams exhibit a typically "Alpine", "glacial" (Pardé, 1933; Godefroy, 1948) or "nival" (Blanchard, 1947) régime, in which maximum summer flow results from the spring and summer melting of snow and ice. Discharge is very low during the winter, falling to as little as 0.03 m^3/s but only occasionally has no discharge been recorded. Winter discharge is derived chiefly from basal ice melting resulting from the geothermal heat gradient (Stenborg, 1965; Vivian, 1967a, 1970c), friction from ice flow (Vivian and Zumstein,1973) and groundwater flow ("fossil" water, Vivian, 1970c)(and see Churski, 1973). Other workers suggest that delayed summer run-off is also an important source of winter meltwater, especially in the larger glacier basins (e.g. Thorarinsson, 1939; Meier and Tangborn, 1961). However, peak discharges occur on the same days in meltwater issuing from each glacier, so that delay in run-off cannot be very severe. Stenborg (1965) claims that the degree of basal melting increases as the extent of the glaciated area increases. He also states that basal melting partly depends on ice-marginal temperatures. Parts of the névé areas of these glaciers may, for example, be frozen to the bed in some winters, thus reducing the amounts of meltwater discharge at that time of year.

Discharge quickly rises to a peak between May and July; an absolute maximum value of 85 m^3/s was recorded at the Mer de Glace in 1966. Discharge declines slowly during September and October, to reach low winter discharge by the end of October or beginning of November (figure 2.14, page 86). The rate at which discharge values rise in spring is slightly sharper than the rate at which they fall in late summer. The initial rise of discharge results from snowmelt, which commences before ice-melt (Pardé, 1933; Godefroy, 1948; Pissart, 1967), especially near exposed rock or where covered with aeolian sediments, and during periods of rainfall, rather than snowfall. Rain acts both as a source of additional run-off and as an agent for melting ice. After several weeks, the rate of melting is accelerated as snow melting is extended and as ice begins to melt. By mid-July, discharge peaks result mainly from ice-melt, accentuated during wet conditions by storm run-off. The area and quantity of snow and ice available for melting are reduced by September, as temperatures begin to fall and the freezing limit is reduced in altitude. The gradual decrease of discharge during September and October is only interrupted by episodes of warm or wet weather conditions.

Daily discharge values of the River Arve, the meltwater stream of the Mer de Glace, for 1970, have been plotted in figure 2.15 (page 87), as this is the most complete set of recent data available. The frequency of extreme peak flows can be determined by attributing probabilities to the maximum annual discharge values (e.g. Pardé, 1933) and plotting on a Gumbel frequency graph (figure 2.16, page 86). The recurrence interval, ri, of peak flows is calculated from $ri = (n+1)/r$, where n is the total number of annual peak flows and r is the ranking order of each peak flow (e.g. Morisawa, 1968; Riggs, 1968; Dury, 1969).

Figure 2.14 Annual regime of meltwaters from five Mont Blanc glaciers

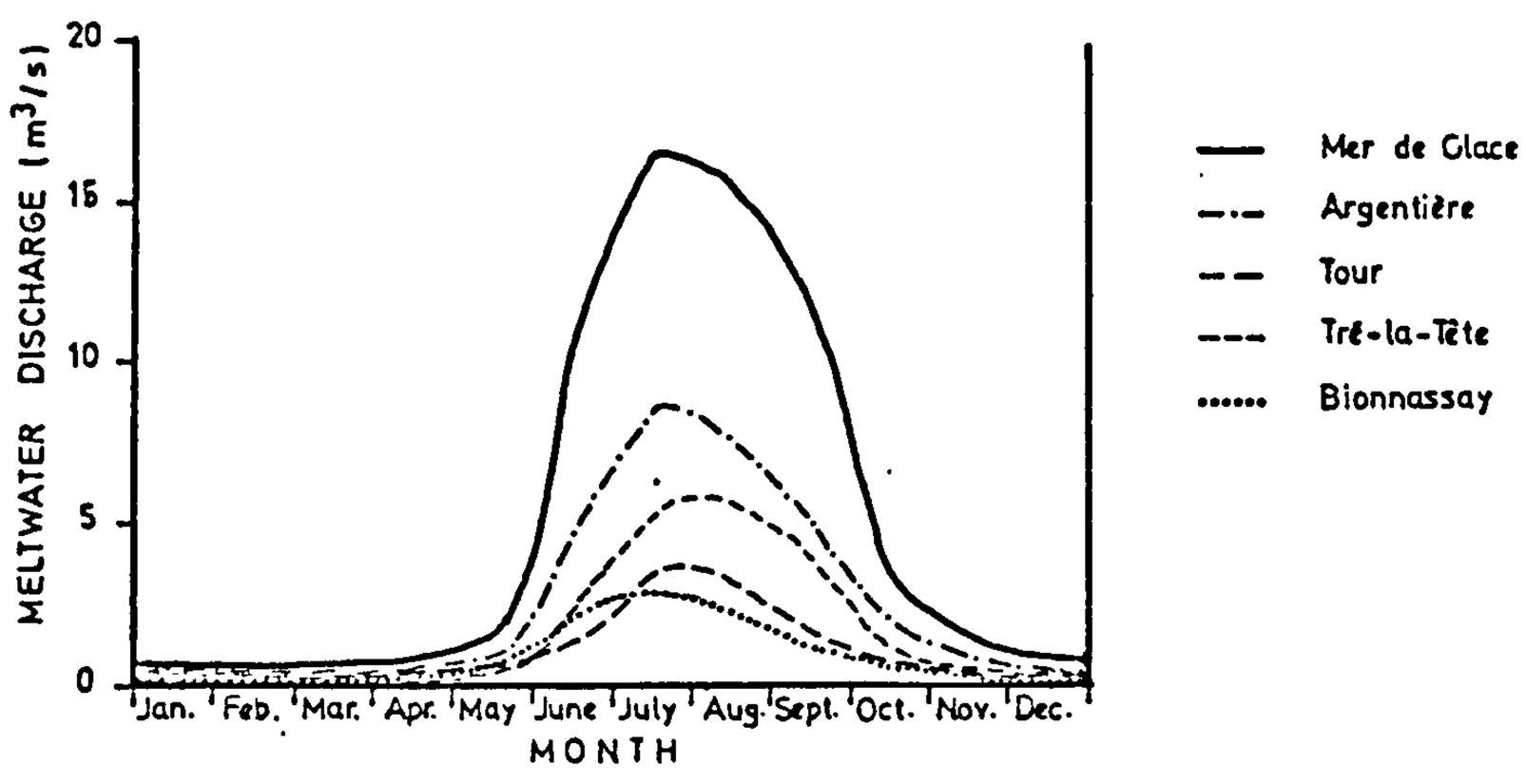
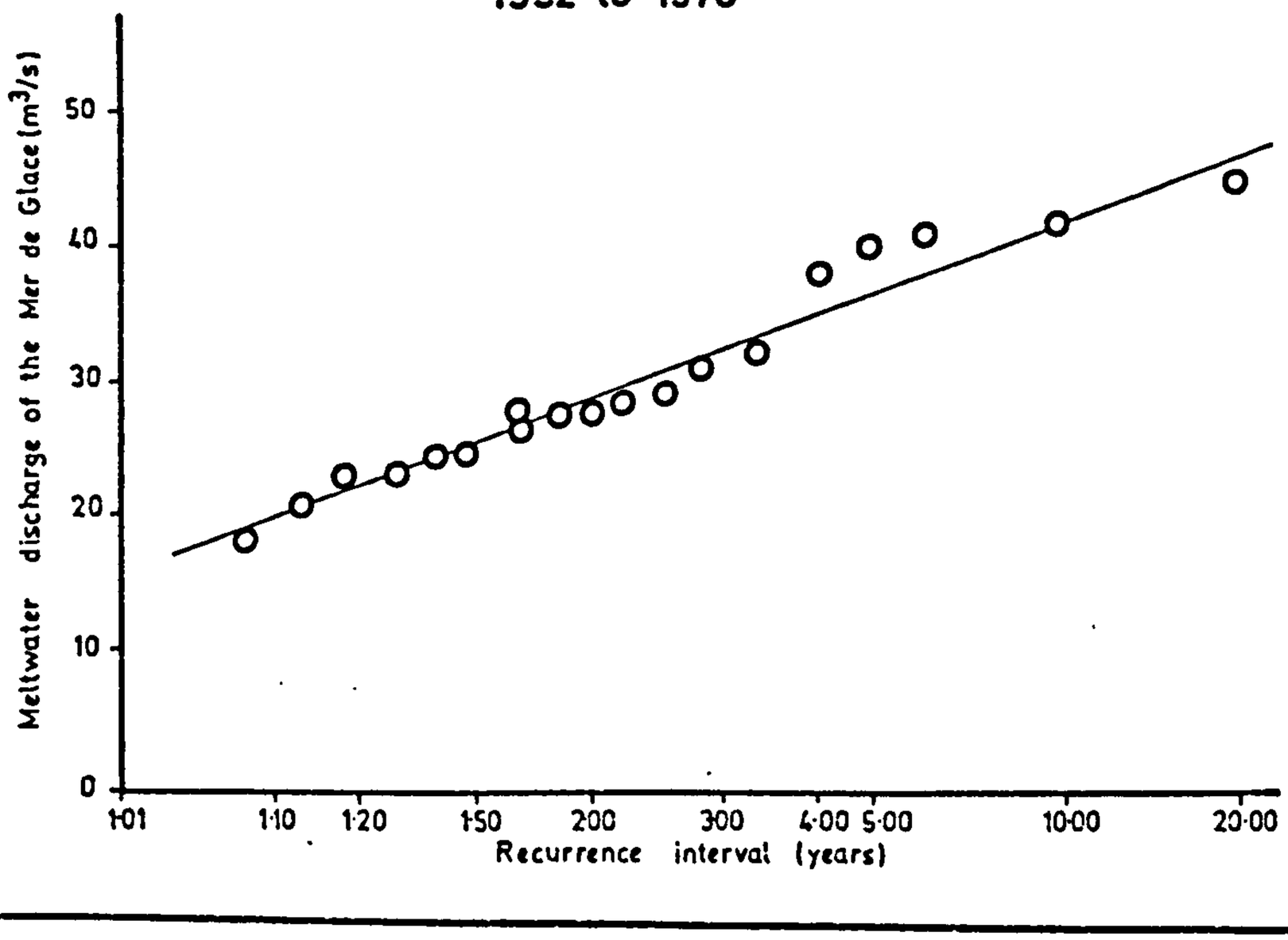


Figure 2.16 Frequency of peak annual meltwater discharges of the Mer de Glace, 1952 to 1970



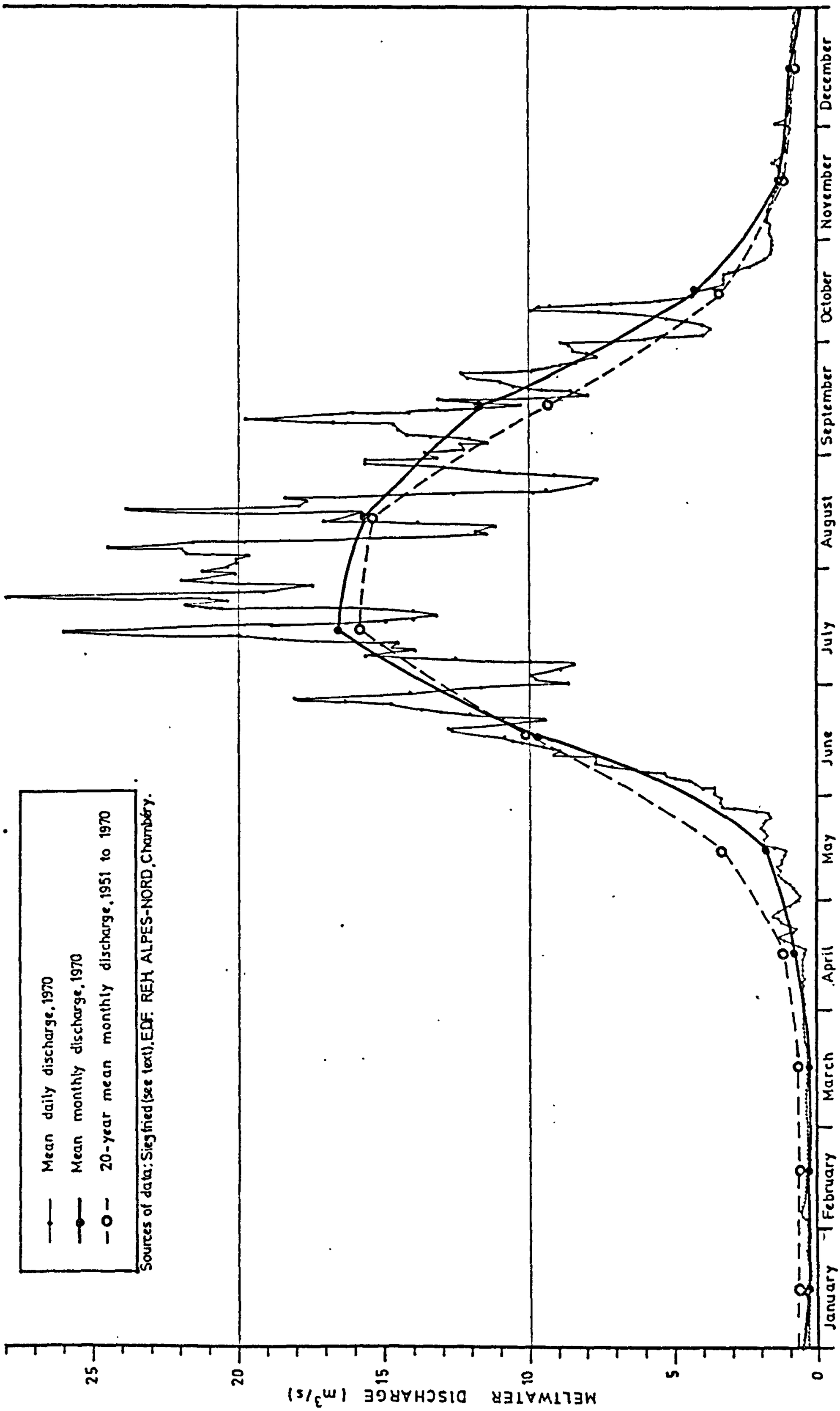


Figure 2.15 Mean daily discharge of meltwaters from the Mer de Glace, 1970

2.5.1.3. The length of the 'meltseason'

An objective method of estimating the 'beginning' and 'end' of the meltseason was sought. It was found that discharge amounts falling below minus one standard deviation from the annual mean discharge corresponded well with the junction between low winter discharge and high summer discharges. The date each year on which the daily mean discharge of each meltwater stream reached the 'critical value' in spring and fell below the 'critical value' in autumn gives an estimate of the length of the meltseason (table 2.5, page 89 , and figure 2.17, page 90).

The beginning and end of the meltseason, as defined above, does not always occur simultaneously in the different glacier meltwaters. Weather conditions within the massif may, therefore, be as important in affecting meltwater discharge at each end of the meltwater season as are internal or local glacier conditions (but see Østrem and Ziegler, 1974). Local conditions may affect run-off after the sudden melting of an iced-up lake, the bursting of an ice-barred englacial reservoir (e.g. Meier and Tangborn, 1961; Bradley et al., 1972; Church, 1972), mass movement of snow from valley-side slopes into the meltwater streams, or storms which cause the critical discharge to be reached in some areas and not in others. The delay in run-off from larger glacier basins appears to be extended over a longer period than from smaller glacier basins (Lütschg, in Stenborg, 1965), but the initial storm run-off is recorded more or less simultaneously. A similar amount of precipitation falling in all areas will raise low discharge values by proportionately larger amounts than it will raise high discharge values. The critical value may therefore be more easily reached by storm precipitation falling over small glacier basins producing low meltwater discharges. This fact is also reflected in the greater constancy of the dates marking the limits of the meltseason from year to year in the larger glacier basins of the Mer de Glace and the glacier d'Argentière, than in the smaller basins of the Tré-la-Tête and Bionnassay glaciers. Winter storm run-off may also be more marked in these two smaller

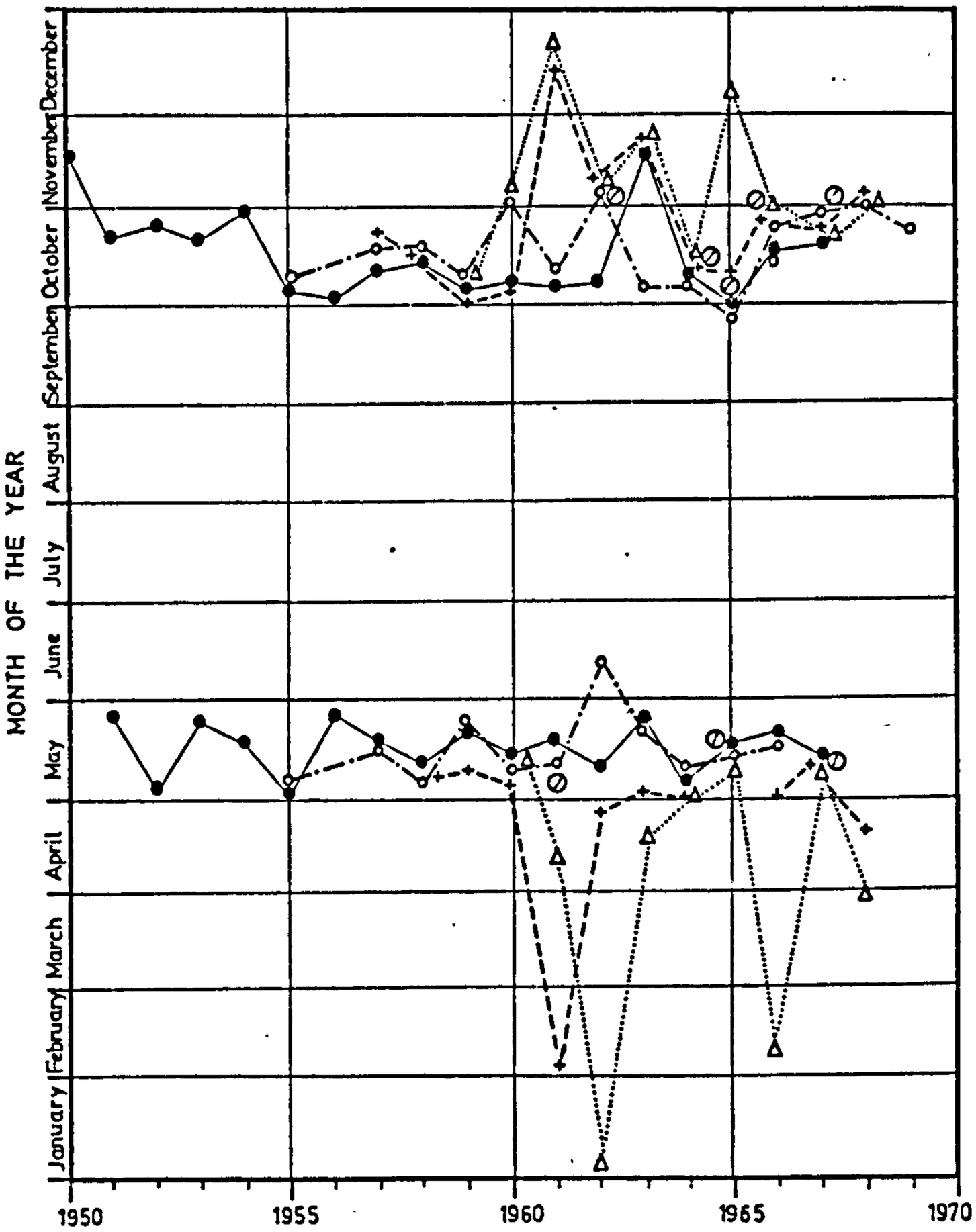
Table 2.5

The length of the meltseason recorded in meltwaters of five Mont Blanc glaciers

Glacier	Bionnassay		Le Tour		Tré-la-Tête		Argentière		Mer de Glace	
Glaciated area (km^2)	3.38		7.75		9.13		16.50		33.44	
'critical discharge value' (m^3/sec)	0.563		0.428		1.018		1.500		3.810	
Year	dates*	No.of days	date	No.of days	dates	No.of days	dates	No.of days	dates	No.of days
1950									18/11	
1951									25/5-22/10	150
1952									2/5-26/10	177
1953									23/5-22/10	152
1954									17/5-31/10	167
1955							1/5- 6/10	158	30/4- 5/10	158
1956							5/10		27/5- 4/10	130
1957					20/10		16/5-20/10	157	19/5-12/10	146
1958					10/5-17/10	160	6/5-17/10	164	10/5-15/10	158
1959	2/10				8/5-11/10	156	22/5- 7/10	138	21/5- 6/10	138
1960	12/5- 2/11	174			6/5- 7/10	154	8/5- 3/11	179	14/5- 8/10	147
1961	10/4-23/12	257	5/5		5/2-12/12	310	9/5-12/10	156	17/5- 7/10	143
1962	2/1- 8/11	301	8/11		25/4- 8/11	197	11/6- 8/11	155	9/5- 8/10	152
1963	17/4-21/11	218			2/5-20/11	202	23/5- 7/10	137	27/5-15/11	172
1964	4/5-11/10	160	8/10		5/5-10/10	158	9/5- 9/10	153	5/5- 9/10	157
1965	14/5- 6/12	206	14/5-6/10	145	9/10		15/5-25/9	133	16/5- 1/10	138
1966	9/2-28/10	261	25/10		1/5-26/10	178	17/5-26/10	162	21/5-17/10	149
1967	12/5-18/10	159	12/5-24/10	165	11/5-20/10	162	29/10		13/5-20/10	160
1968	30/3- 3/11	218			21/4- 3/11	196	28/4- 3/11	189		
1969							24/10			
mean** (rounded values)	3/4- 6/11	217	10/5-19/10	155	25/4-26/10	187	14/5-18/10	157	17/5-15/10	152
standard deviation** (days)	57-50	91	29-47	156	47-41	78	42-32	50	36-29	39

*12/5-2/11 indicates that the 'meltseason' commenced 12 May and ended 2 November.
**Two sets of figures relate to the beginning and end of the 'meltseason' respectively.

Figure 2.17 The beginning and end of the meltseason of five Mont Blanc glaciers, 1950 to 1969



glaciers because they lie further west and receive marginally greater amounts of precipitation.

The average date of the beginning of the meltseason ranges from the 3rd April (Bionnassay glacier) to the 15th May (Mer de Glace), a difference of about six weeks (see Golubev, 1973; Pinchak, 1973). The meltseason thus begins in early April, but, for large basins, it takes about six weeks for melting to extend over the whole basin to produce sufficiently high discharge amounts to reach the critical value. The average date of the end of the meltseason ranges from the 17th October (Mer de Glace) to the 6th November (Bionnassay), a difference of about three weeks. This suggests that the smaller glaciers respond markedly to storm precipitation later in the year. The greater time difference in spring between the beginning of the meltseason in different basins, compared with the end of the meltseason in autumn, reflects the large proportion of spring meltwater which is derived from snowmelt and the rapid response of small glaciers to storm run-off.

The smaller glaciers, as a result, can have a significantly longer 'meltseason' than the larger glaciers, although variability in the length of the meltseason is much greater (figure 2.18, and table 2.5). The 'meltseason' at the larger glaciers thus depends more on changes in air temperature, which are relatively uniform from year to year, than on spasmodic periods of precipitation.

The smaller glaciers not only have a longer and more variable meltseason, but despite a greater melting period, produce much smaller quantities of meltwater. The glaciated area[1] of the five main glaciers is closely correlated with mean monthly meltwater discharge amounts (figure 2.19, page 93) (see also Golubev, 1973; Kasser, 1973).

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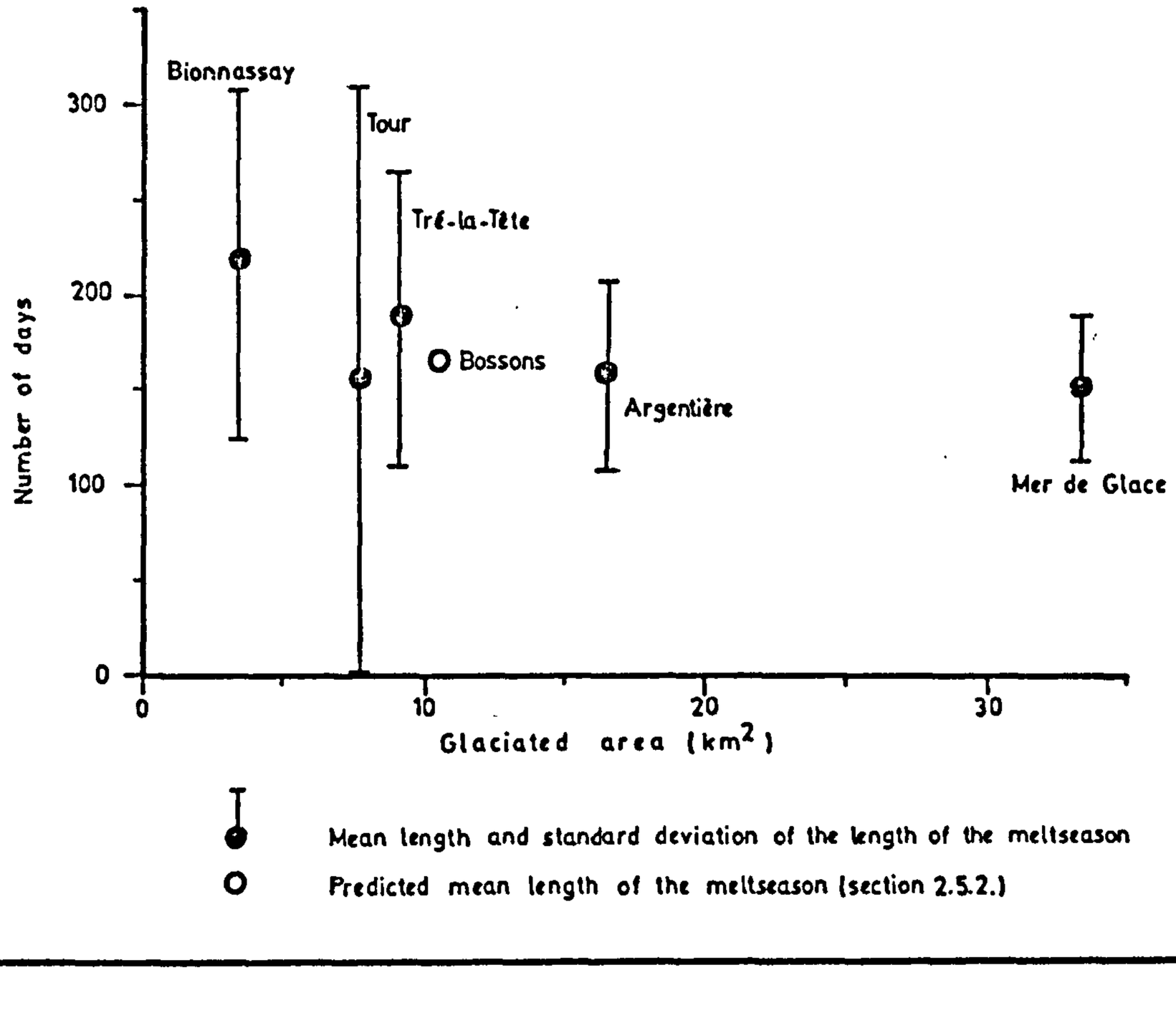
1. The glaciated area is that area measured, in km^2, as ice-covered, from the 1959 topographic maps (I.G.N.) and measurements by Vivian (1967b,c, 1969a,b, 1970a,b).

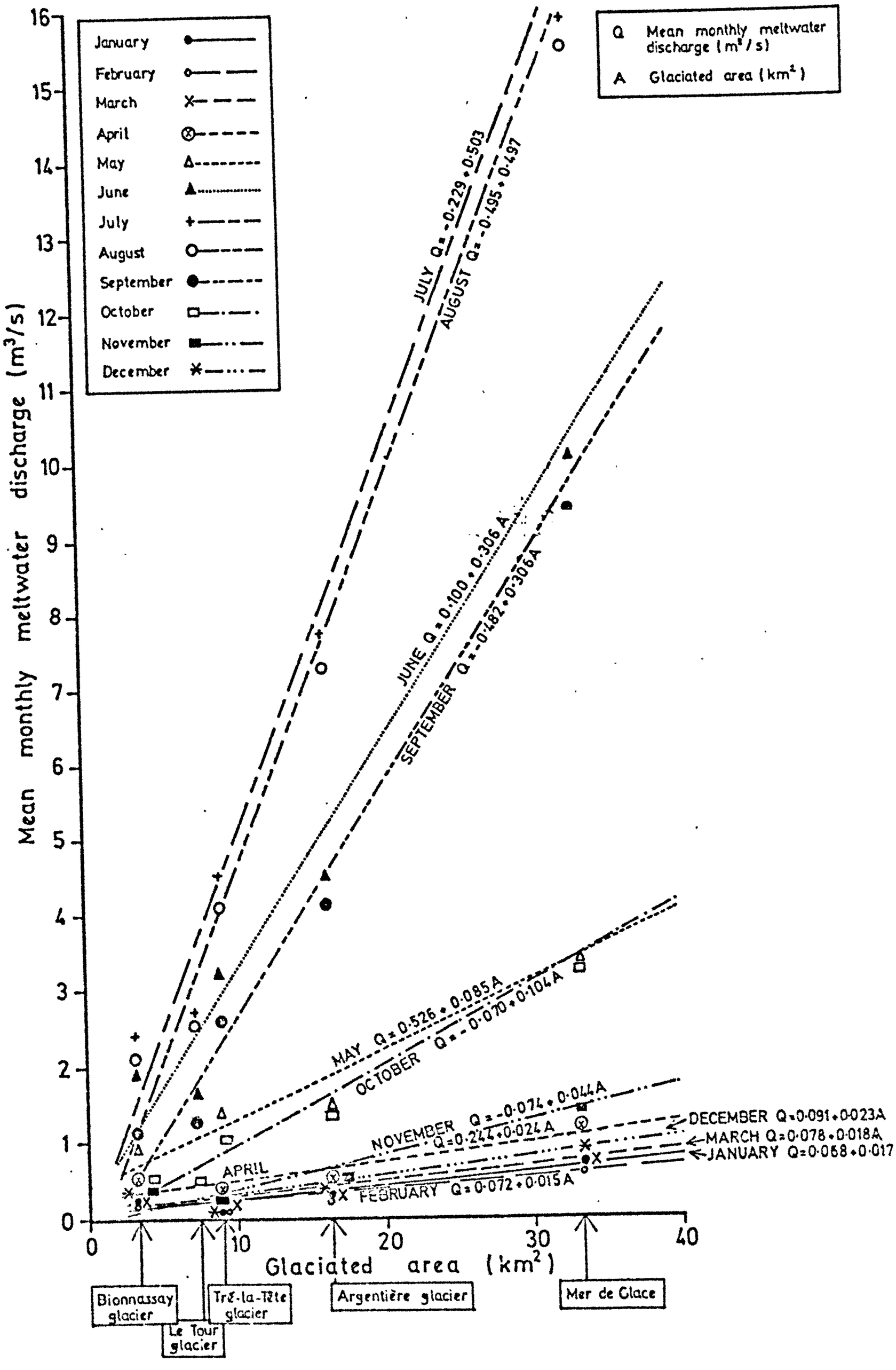
Figure 2.18 The length of the meltseason in six glaciated basins in the Mont Blanc massif



The high correlation between glaciated area and meltwater discharge at all times of the year within the massif, suggests that the whole glacier basin contributes meltwater both in summer (from surface and basal ice-melting, and from precipitation run-off) and in winter (from basal ice-melting, precipitation run-off, and delayed summer run-off). A relationship between basin area and discharge is one which has been widely accepted in non-glaciated fluvial basins (e.g. Hack, 1957; Hack and Goodlett, 1960; Brush, 1961) and it appears that the same kind of relationship may also hold good in glaciated basins which are located in similar terrain. Stenborg (1965) also states that the amount of run-off is at least partly dependent on glaciated area.

Important seasonal variations in this relationship have emerged. Firstly, correlations, although still significant at the 0.05 significance level, are lower in winter, for winter discharge results largely from variable localized, internal glacier conditions, whilst summer discharge results mainly from external climatic conditions.

FIGURE 2.19 MONTHLY MELTWATER DISCHARGE OF FIVE GLACIERS OF DIFFERENT AREA IN THE MONT BLANC MASSIF



Secondly, the gradient of the best-fit line (the 'beta'-coefficient) increases in steepness between February and July, and then decreases from August to February (see figure 2.19). This means that the difference in meltwater discharge from glaciers of different sizes is much less in winter than in summer. Winter discharge is relatively small, even from the larger glaciers. The differential between the total amount of melting between the small and large glaciers increases as spring and summer temperatures become more effective in causing ablation, and as the whole glacier begins to contribute meltwater.

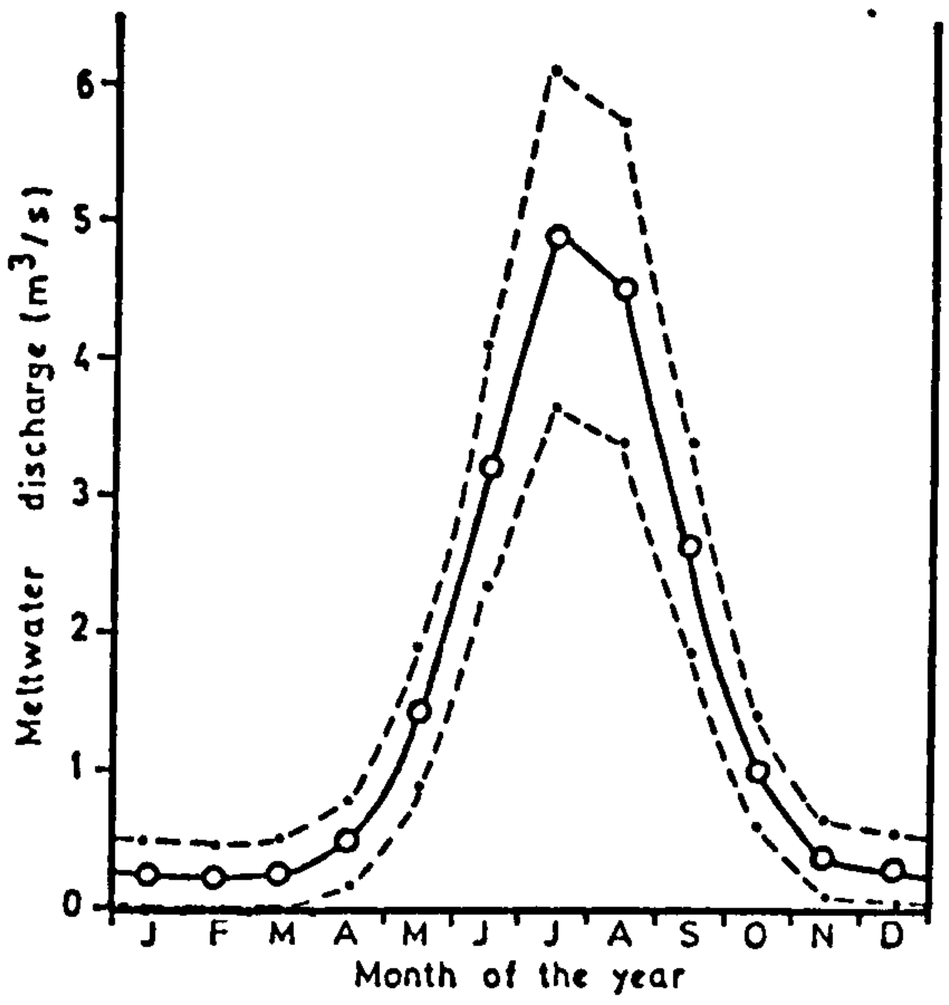
2.5.2. THE BOSSONS GLACIER

2.5.2.1. The annual régime

The lack of any long-term gauge recordings of meltwater discharge from the Bossons glacier has necessitated the estimation of the annual hydrological régime. The estimates are based on (1) the relationship between glaciated area and meltwater discharge which was found to exist for the five neighbouring glaciers (figure 2.19) and (2) the calculated relationship between mean monthly discharge and each respective standard error of discharge, and thence the probability of flows of different magnitudes. The first set of estimates is based on the glaciated area of the Bossons glacier, assuming that the overall relationship can also be applied as a result of regional homogeneity of response to climate per ice-covered areal unit.

The resulting curve (figure 2.20, page 95) exhibits an annual pattern similar to the other meltwater streams. The 'meltseason' begins at the end of April and lasts until early October, extending over a period of $5\frac{1}{2}$ months (see figure 2.18, page 92). 'Peak flow' occurs between mid-June and late August, a period of 2 to $2\frac{1}{2}$ months. The average flow in February (month of minimum flow) is about $0.23 m^3/s$ compared with about $4.9 m^3/s$ in July. Mean

Figure 2.20 Predicted monthly meltwater discharge of the Bossons glacier



summer discharge is thus over twenty times greater than mean winter discharge. This compares well with Pardé's statement (1933) that summer meltwater discharge from glaciers is between 15 and 20 times greater than winter discharge, with Corbel's (1963) corresponding discharge values of between 4 and 25 times, and with Tricart (1970) who found summer discharges to be 10 to 20 times greater than winter discharges.

The Bossons glacier is, however, drained by three major meltwater outlets. One outlet is from the ice-snout on to the proglacial valley-train (at an altitude of 1230 m), while the other two leave the eastern part ice-margin above 2800 m and flow downslope through deep gullies (see figure 2.11, page 74). Measured data suggest that discharge amounts in the valley-train are only about one eighth of the total meltwater discharge (see below).

2.5.2.2. Direct measurements of meltwater discharge

Direct measurements of meltwater discharge were carried out during April and July 1971. They comprise short-term, discontinuous measurements of discharge based on hourly recordings (during daylight hours) of the width, depth and surface velocity of three meltwater stream reaches on the valley-train. Reaches 'B' and 'C' were on two large meltwater streams emerging from either side of the ice-snout. Reach 'A' was on the only outlet stream from the valley-train (see figure 2.5, page Had there been a single meltwater stream emerging from the ice-snout, as in 1969, the amount of input and output of water and water-carried sediment could have been determined.

The width and depths of the cross-profile at the downstream end of each reach were measured, and the average depth calculated. Surface velocity was measured by timing the rate of movement of coloured dye[1] from the upstream to the downstream end of the reach. This process was repeated five times and the average surface velocity calculated. Mean velocity was taken as 0.85 that of the surface velocity (Østrem, 1964), from which discharge was calculated in m^3/s (see Appendix 1).

Discharge values were of a very low order. At section A, discharge ranged from only 0.120 to 0.260m^3/s in late April to 0.400 to 0.616m^3/s in mid-July. Measured discharges in July are less than five times greater than in April. The total discharge is very small at all times, and probably less than the 'predicted' values. However, mean monthly values, which formed the basis of predictions, cannot be calculated using the present data.

Large diurnal variations in discharge were recorded, especially at the height of the meltseason. Many other workers have also recorded diurnal discharge variations (e.g.

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1. Depths were often too shallow to allow free movement of any floating objects (e.g. plastic floats). Dye velocities at each of the five measurements per hourly recording were always consistent.

Meier and Tangborn, 1961; Liestøl, 1967; and Tricart, 1970; and see chapter 5). Morning discharge decreased until about 1100 hours, subsequently rising rapidly to a peak in late afternoon, between 1700 and 1800 hours. Insolation is sufficient by mid-day to cause rapid and extensive ice-melting, especially in marginal areas. During the late afternoon as temperatures cool down, discharge still remains fairly high. Although less ice is melting at that time, there is much residual meltwater arriving from more distant parts of the glacier. In a very large glacier basin, no diurnal discharge variation may even be apparent (e.g. see also Stenborg, 1970). The supply of residual meltwaters from the Bossons glacier becomes exhausted by early the following morning, by which time discharge is approaching a minimum. The minimum will occur much earlier if freezing occurs overnight.

Major discharge peaks occurred after summer storms, while minor peaks of flow occurred during hot, dry weather. Many workers have also considered the relationship between discharge and weather conditions, and have come to similar conclusions (e.g. Keller, 1916; Rogstad, 1951; Arnborg, 1955; Østrem et al., 1967; Church, 1972).

2.6. SOILS AND VEGETATION

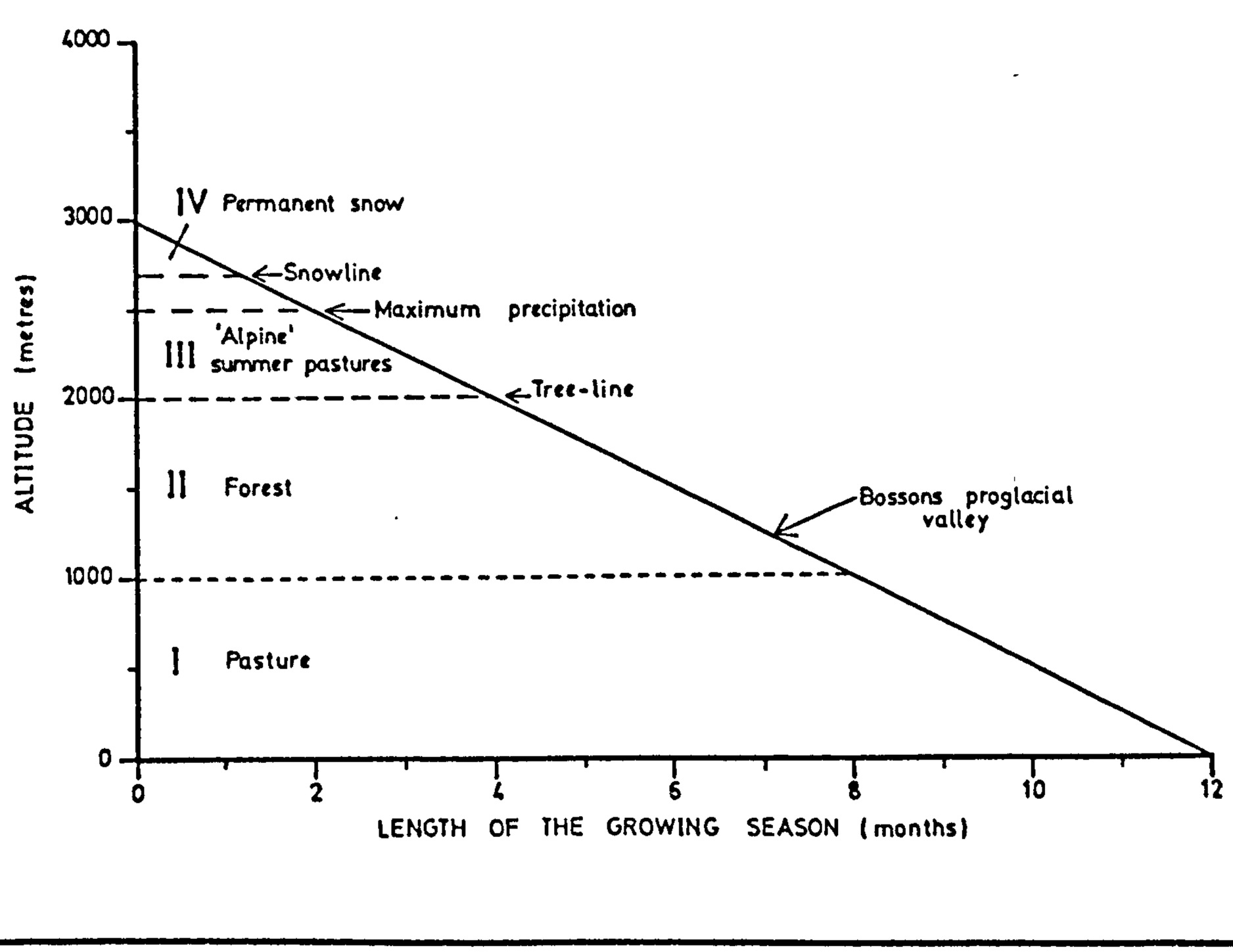
2.6.1. THE MONT BLANC MASSIF

The literature provides a general outline of vegetation distribution in the massif, and the characteristics of the vegetation which inhabits the mountain environment (e.g. Université de Grenoble, 1963, in the French Alps; Leach (1930) and McVean and Ratcliffe (1962) in Britain; Dahl (1951) in Scandinavia; and summaries in Blanchard (1958) and Godefroy (1948) of plant environments in the western Alps). Few works have been published on Alpine soil types (e.g. Retzer, 1956; Cooke and Lawrence, 1959; Bouma and Plas, 1971).

2.6.1.1. Altitudinal vegetation zones

The length of the growing season, defined as the period during which mean monthly temperatures rise above 6-10°C (Griggs, 1937), decreases with increasing altitude, and the type of vegetation varies accordingly. Forest growth requires a growing season of over four months, and in the Western Alps this is only possible at altitudes of less than about 2000 m. The upper limit of forest growth, marked by the 'treeline', does, however, vary locally in altitude depending on aspect, exposure, slope, drainage and soil development. Godefroy (1948) outlines four altitudinal vegetation zones based mainly on the length of the growing season and the altitude of the permanent snowline (figure 2.21).

Figure 2.21
Altitudinal vegetation zones in the Western Alps
(After Godefroy, 1948)



2.6.1.2. The Alpine plant environment

The type of vegetative growth depends on edaphic features such as temperature, quantity and quality of light, wind force, humidity, parent material, degree of rock weathering and extent of soil development. Edaphic conditions in an Alpine environment have some of the following characteristics:

(1) Strong light intensity, especially ultraviolet radiation, which increases the rate of plant growth so that the annual growth cycle may be completed in the shorter growing seasons which occur at high altitudes;

(2) High summer temperatures (rising to 20°C) but low winter temperatures (4 to 5 months below 0°C) and a large diurnal temperature range;

(3) Much precipitation (generally over 1100 mm), with a large proportion falling as snow;

(4) Strong and persistent winds, which increase the rate of evapo-transpiration and encourage plant colonisation to occur in sheltered localities;

(5) Unstable soil mantle, resulting from both steep slopes and frequent ground frost activity; and

(6) Soils that are dry, or effectively dry when frozen, with little moisture available from subsoil for plants.

These features of the Alpine environment are superimposed on local conditions, which therefore provide varied habitats for vegetation. In addition, since each plant has unique limits of tolerance of physical conditions, vegetation distribution is locally quite varied.

2.6.2. THE BOSSONS VALLEY

The proglacial valley of the Bossons glacier lies about 800 m below the local treeline. Vegetation development on deposits exposed by glacier retreat therefore follows a succession from mosses and lichens, to 'pioneer' plants, shrubs and trees, and finally to the 'vegetation

climax' of spruce (Picea abies) forest. The Bossons valley, lying at an altitude of about 1200 m, should experience an average growing season of just over 7 months from April to October (see figure 2.21), but the glacial influence on the microclimate may reduce this period in some parts of the valley (section 2.3.).

2.6.2.1. Weathering and soil development

Exposure to the atmosphere of deposits uncovered by retreating ice leads to the gradual modification of these mineral-rich deposits to form well-humified, vegetated soils (Crocker and Dickson, 1957; Crocker and Major, 1955; Stork, 1963). One of the most important initial soil-forming processes is the leaching out of soluble elements such as calcium (e.g. Leighton and MacClintock, 1930; Reiche, 1945). The progressive removal of calcium eventually has a measurable effect in increasing the acidity of the substratum.

In order to determine the degree of leaching which had occurred in proglacial deposits of different origins, the pH values of 193 sediment samples were measured (figure 2.22, page 101).[1] The samples were selected randomly from the valley-train, and systematically from the two lateral moraines and the terminal moraine. In most cases, the substratum was unvegetated. The results have been discussed in detail in an earlier publication (Maizels, 1973).

The 'western' lateral moraine is composed predominantly of calcareous and crystalline schists (section 2.1.). The pH of morainic debris close to the ice-margin ranged from 7.5 to 8.0, i.e. very alkaline, and gradually decreased away from the glacier to 6.0 to 6.5 at 230 m from the ice-margin (figure 2.22b, page 101). The 'eastern' lateral moraine, by contrast, is largely composed of 'protogine' granites. Morainic deposits close to the ice-margin are acid (pH = 6.5). Acidity decreases only slightly

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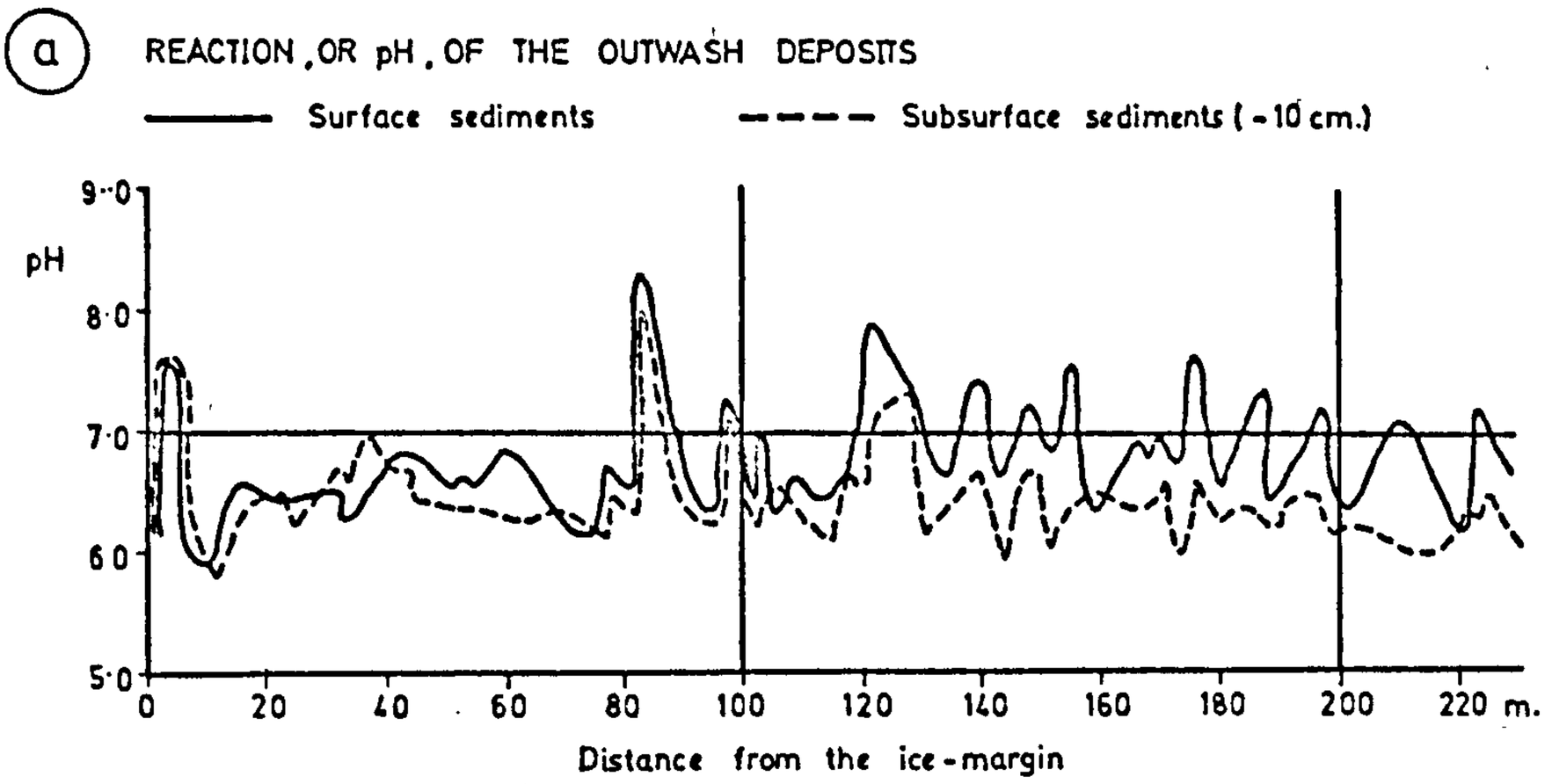
1. A Pye-pH meter was kindly provided by Wye College, University of London.

Figure 2.22
Acidity of the Bossons proglacial deposits



Based on Maizels, 1973

down-valley for a distance of 180 m, after which vegetation growth appears to have reduced the acidity to 6.1 (figure 2.22b). The transition between the alkaline western moraine and the acid eastern moraine is most clearly seen in pH values of samples across the terminal moraine (figure 2.22c).

Persson (1964) and Viereck (1966), working on Icelandic glacier forelands, also recorded increases in acidity with distance from the ice-margin. pH values decreased from 8.4-8.0 to 5.3-4.4, after exposure lasting up to 300 years. Stork (1963) measured a pH of 8.1 at the snout of a Norwegian glacier, situated on basic amphibole bedrock, and acidity increased downvalley to a pH value of 5.5. The increase of acidity downvalley thus reflects the gradual retreat of the ice-margin, exposing downvalley locations to subaerial leaching.

The pH values of the fluvioglacial outwash sediments of the Bossons glacier, however, fluctuate continuously downvalley (figure 2.22a, page 101). Subsurface samples, taken from 10 cm below the surface, were more acid than their equivalents on the surface, having a mean pH value of 6.41 compared with 6.73 on the surface. Greater acidity below the surface probably results from increased saturation by acid meltwaters, derived especially from snowmelt (Clement and Vaudour, 1967), which seeps underground. During the meltseason, the watertable lies less than 10-15 cm below the surface and, in addition, groundwater flow tends to remove silts and clays (Chapter 6), which might otherwise act as a buffer to the lowering of pH (Viereck, 1966). Trend surface analysis of the acidity of subsurface outwash deposits (figure 2.23, page 103) reflects the higher alkalinity of sites close to the ice-margin, despite relatively low degree of statistical explanation.

The degree of acidity is therefore locally dependent on the composition of the sediment, the frequency and degree of saturation and the origin of the floodwater. The wide variability in acidity of the outwash deposits, compared

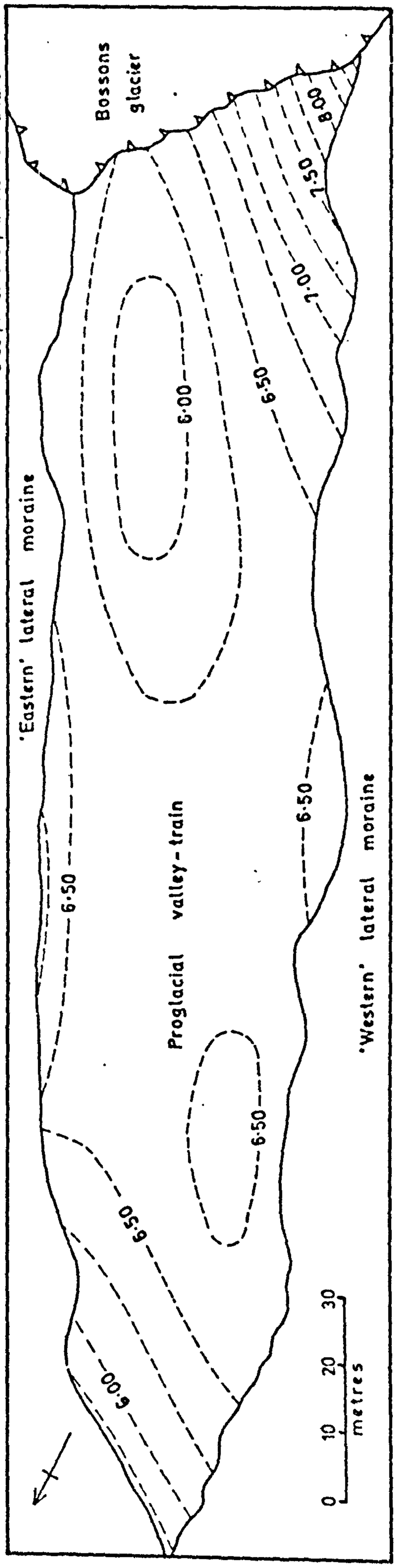


Figure 2.23
Cubic trend surface of acidity of subsurface outwash deposits, Bossons valley-train, 1971

with the systematic changes recorded on the moraines, indicates that the valley-train sediments are too unstable to allow gradual leaching over time.

On the lateral and former terminal moraines, however, soil development has progressed, as indicated by the change in pH, to a suitably acid substratum, and vegetation has successfully colonized many areas (figure 2.24, page 105).

2.6.2.2. The development of vegetation

Vegetation development has occurred despite proximity to the ice, which modifies proglacial temperatures and air currents. Precipitation and meltwater provide a fairly adequate water-supply for plants, although periodic flooding and continual substratum instability provide adverse conditions for plant colonisation in the valley.

There are four plant 'habitats' within the proglacial valley, and the substratum of each of these habitats is composed of high proportions of stones and coarse gravel (Chapter 6).

A. The 'western' lateral moraine is an unstable, schistose debris slope (30-33°), and is effectively a 'ubac' slope (figure 2.7, page

B. The 'eastern' lateral moraine is also an unconsolidated debris slope (22-27°), containing a high proportion of protogine boulders. This is an 'adret' slope. Both lateral moraines are well-drained and often dry.

C. Former terminal moraine, forming a central 'island', is only 12 m in length (in 1974), is being progressively buried by meltwater flood deposits on the valley-train, and is under constant and direct attack by cold air currents blowing off the glacier.

D. Proglacial outwash deposits are everywhere moist, and most areas are flooded during periods of high

Figure 2.24 Extent of vegetation development in the Bossons proglacial valley

Figure 2.25 Pioneer vegetation colonisation on coarse outwash of the Bossons valley-train

meltwater discharge. Floods remove some fine sediments, while some filter downwards, but 'silt-skins' are redeposited on the surface during the waning of a flood. This habitat is much more level, and the instability of the substratum reflects powerful meltwater flow rather than gravity action which operates on the steeper slopes.

A plant succession observed in these areas of the proglacial valley has followed three main stages.

I. Primary colonisation of bare rock and gravel surfaces by pioneer plant species is taking place in proximal areas of habitats A and B, and in raised and abandoned tracts of habitat D (figure 2.25, page 105). Pioneer plants have long, binding roots and encourage the accumulation of litter which, with the mineral content of the substratum, will form the soil. Pioneer plants are frequently found in the lee of a boulder or in a hollow, where a pocket of soil has developed. The most extensive pioneer species was Epilobium Dodonaci, while on the more stable banks of outwash Trifolium repens, Saxifraga aizoides and Agrostis stolonifera had become established. Stage I colonisation may take place only 1 or 2 years after exposure from ice.

II. New plant species, trees and shrubs migrate into the valley after five to ten years, to fill new niches on the more mature soils. Dense thickets of alder and willow species have become extensive over the distal slopes of habitats A and B, and cover the former terminal moraine, habitat C (figure 2.24, page 105)(and see Chapter 3). The main tree species include Betula pendula, B. incana, Salix aurita and S. cinerea. Some occasional individuals of Larix decidua have also become established.

III. Picea abies becomes more important after 50-60 years, and after 100-120 years it is the dominant tree species, representing the local 'climax' vegetation. Picea abies was found only at the distal end of the 'eastern' moraine, habitat B.

Similar plant successions in front of glacier terminii have been recorded by other workers. These include Girardin (1902) and Lüdi (1958) in the Alps, Chouard (1933) in the Pyrenees, Hale (1952) in Baffin Island, Stork (1963) in Norway, Persson (1964), Lindroth (1965) and Wojcik (1973a) in Iceland, and Viereck (1966) in Alaska.

The high alkalinity of many parts of the valley is also illustrated by the presence of calcicole plants (base status over 30mg/100gm and pH over 4.8). Calcicole indicator plants included Anthyllis vulneraria, Crepis, Fragaria vesca, Geranium robertianum, Trifolium repens, Tussilago farfara, and Saxifraga aizoides. Many plant species are also indicators of damp substratum conditions prevalent on the valley-train and in areas liable to meltwater flooding. These species include Agrostis stolonifera, Alnus incana, Betula pendula, Crepis, Epilobium Dodonaci, Festuca rubra, Parnassus palustris, Salix aurita, Saxifraga aizoides and Tussilago farfara.

A more detailed description of the location of different plant species and their physical requirements is included in Maizels (1970, 1973).

Once vegetation has been established, each year should see an increase in the extent of vegetation cover, vegetation density and number of species (31 species were recorded in 1969). Flooding followed by erosion or burial of pioneer species, and glacier advance over vegetation in the Bossons proglacial valley have, however, destroyed areas of vegetation (figures 2.26 and 2.27, page 108). The success of vegetation development elsewhere has resulted mainly from the availability of a stable substratum for colonisation, rather than on reduced glacial influence on the climate with increasing distance from the glacier. Indeed, plant migration is extending from the distal towards the proximal proglacial slopes.

Figure 2.26 Flooding of vegetation by meltwaters

Figure 2.27 Over-riding of vegetation by advancing ice

2.6.2.3. Botanical evidence of the age of moraines

The study of vegetation development affords some clues as to the length of time since different proglacial areas have been freed from ice-cover, thus indicating the age, or minimum age, of moraines. Dendrochronology provides a direct method of estimating the minimum date of exposure from ice-cover of those parts of the proglacial area which are vegetated by trees. Dendrochronological methods have been widely used to aid dating of moraines deposited during historical time (e.g. Lawrence, 1949; Harrison, 1952; Heusser et al., 1954; King, 1959; Sigafoos and Hendricks, 1961; Larmarche and Fritts, 1971). Seven tree cores of Larix decidua (European larch) were taken with a Swedish increment borer (Lawrence, 1950) from the base of the trunk of the largest, or only species within the vegetated stand (figure 2.5, page 62).

The distal terminal moraine, on which grow trees with 21-29 rings, was probably exposed at around 1940, assuming the absence of frost or false rings (Glock and Reed, 1940; Hustich, 1949). This date corresponds to the mapped position of the ice at that time (figure 2.12, page 81). The 'central island' terminal moraine contains Larix decidua trees of 17 and 15 years old, suggesting that the ice-snout retreated from this location in 1949-1950. This would allow up to five years for the successful germination of the tree species, following ice retreat. Other workers have also suggested a similar initial period before the substratum is sufficiently stabilized to allow permanent vegetation growth (e.g. Lüdi, 1958; Sigafoos and Hendricks, 1961; Persson, 1964). Analysis of cores from intervening sites support the evidence of a continuing ice retreat.

Lichens were briefly examined, but they provided no additional evidence of glacier variation and the date of exposure from ice (but see Beschel, 1950, 1958). Rhizocarpon geographicum had been selected as being conspicuous (yellow thalli with black rim), with a suitable growth period and growth rate (Benedict, 1967) and found extensively on

protogine granite boulders. The maximum diameter of Rhizocarpon was found to increase progressively from 2 mm at 335 m to 50 mm at 765 m from the profile R_1-R_1 bis (figure 2.12 , page 81), the latter site lying close to the last maximum glacier extent in 1818. This would suggest a growth rate of 0.291mm/p.a. if a period of fifty years is allowed for initial lichen colonisation in otherwise unstable conditions, as suggested by Benedict (1968), and is almost identical to the 0.3mm/p.a. recorded by Beschel (1950) for the average growth rate of Rhizocarpon in front of 24 glaciers in the eastern Alps (see also Andrews and Webber, 1964; Bogacki, 1973). The largest lichens measured at Bossons date from periods following the advances of 1770 and 1818 (see table 2.2, page 78) and support mapped evidence of ice retreat.

2.6.2.4. Conclusion

The three main proglacial vegetation 'zones' have been mapped (figure 2.29, page 111). The first zone lies within 230 m of the ice-margin and is composed of pioneer vegetation initiated since 1940. The second zone comprises vegetation growth occurring after the glacier advance of the early 1890's. Vegetation in this 'post-1890' zone becomes younger as one approaches the glacier because of the subsequent glacier advance in 1920, into proximal parts of this zone. A distinct 'trimline' (Lawrence, 1949; Sigafoos and Hendricks, 1961) separates the younger vegetation of zone II from the older, taller, mature pine and spruce forests (figure 2.28, page 111). The 1890-trimline is so well-marked (see also figure 2.11, page 74) that it could be used to define the former extent and profile of the glacier tongue (see Heusser et al., 1954). The forested areas beyond the trimline represent the third zone of pre-1890 vegetation. Parts of this zone were first forested after the glacier advances of 1854 and 1818, other parts after the advances of 1723 and 1644, but most areas have probably been forested since before "Little Ice Age" times.

Figure 2.28
Part of the 1890 trimline, Bossons glacier

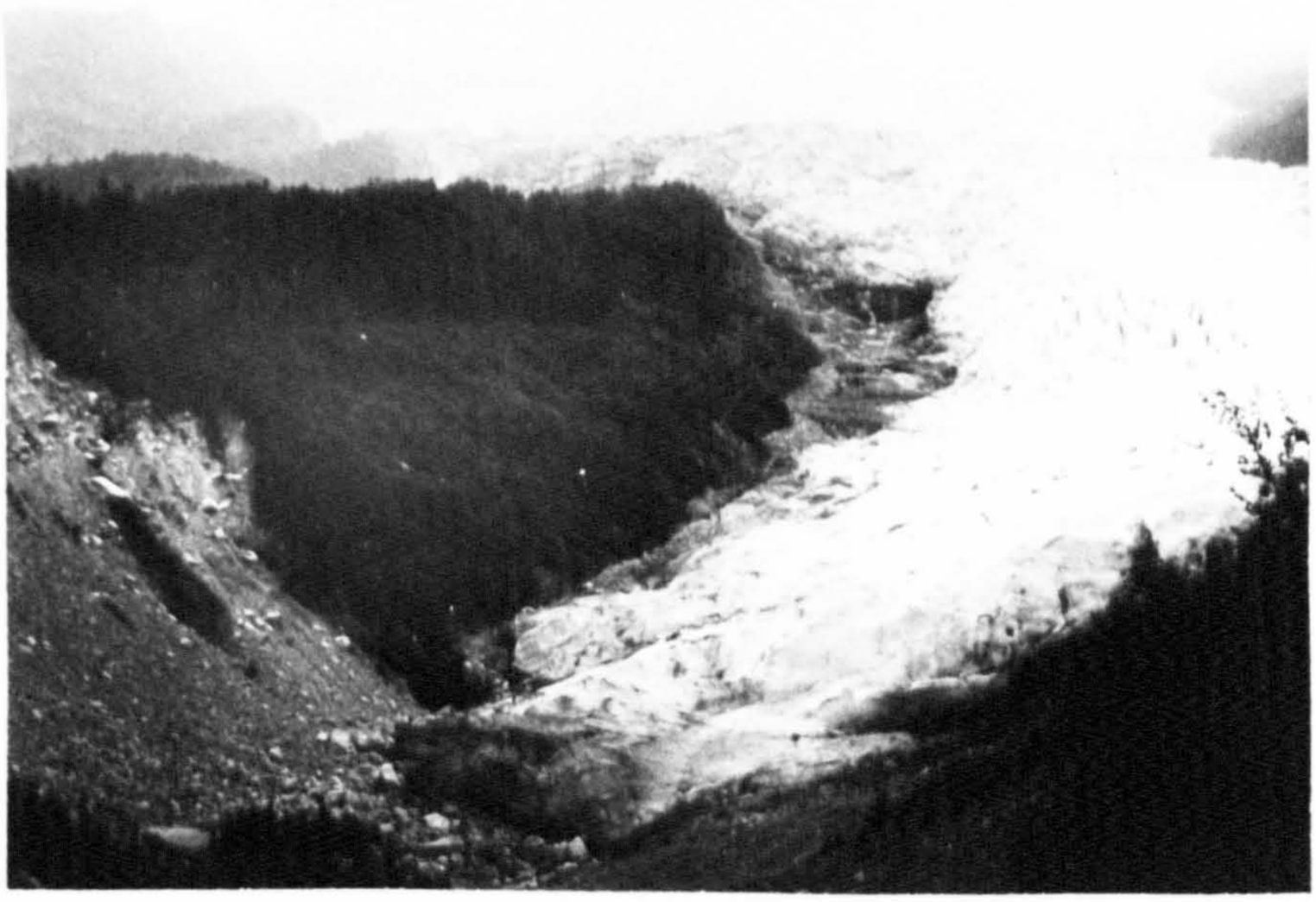
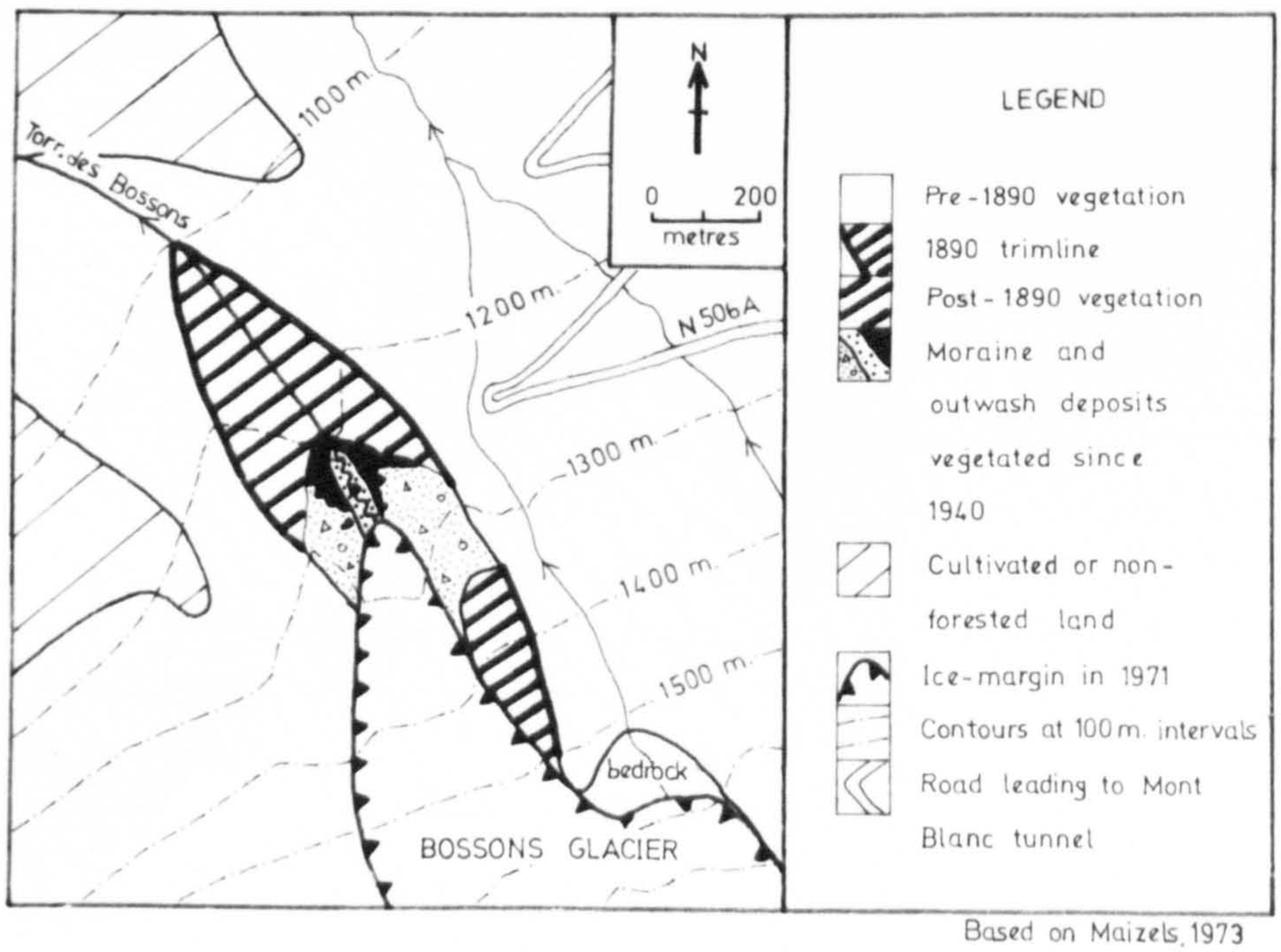


Figure 2.29
Vegetation zones infront of the Bossons glacier



Based on Maizels, 1973

Chapter 3

B O S S O N S II:

P R O G L A C I A L M O R P H O L O G Y

Additional list of symbols

A	Cross-sectional area
a	Intercept coefficient in regression equation
c	Coefficient of proportionality in exponential regression equation
e	Base of natural logarithms
F_R	Graf's form ratio (1970)
k	Permeability of sediment
r	Correlation coefficient
r^2	Coefficient of determination
S	Slope
S_o	Slope at origin, source or ice-margin
SE_b	Standard error of b
s	Slope of water surface
μ	Dynamic viscosity
X	North-south component in trend surface analysis
x	Distance from source or ice-margin, in theoretical regression equations
Y	East-west component in trend surface analysis
y	Dependent variable in theoretical regression equation
Z	Altitude component in trend surface analysis.

Chapter 3

B O S S O N S II: P R O G L A C I A L M O R P H O L O G Y

3.1. THE ICE-MARGIN

3.1.1. THE SHAPE OF THE ICE-SNOUT

The Bossons glacier advanced more than 100m during the period of observation from 1968 to 1974 (figure 3.1, page 115). The rate of advance has varied from place to place along the ice-snout. The eastern section remained in the most advanced position until 1971 as it was able to spread over gentler lateral slopes and extend farther down into the valley than ice in the west, which was obstructed by the high lateral moraine. More recently, ice in the west has been able to advance more rapidly, for, as the ice increased in thickness, it overrode a former debris fan deposit at the foot of the lateral moraine (figure 3.2, page 116). The present ice-snout lies almost perpendicular to the valley axis.

The slope of the ice at the ice-snout may indicate the rate of ice-marginal ablation compared with the rate of ice advance (e.g. Lister, 1958; Price, 1973). A steep ice-cliff can indicate an advancing glacier, in which the advancing ice-mass is too great to be reduced significantly by ablation. A flatter ice terminus results from ablation which is sufficiently effective to melt much of the terminal ice. The snout of the Bossons glacier has been characterized by each of these shapes according to the season, the year and the location on the ice-margin.

(1) Ice-cliffs. The ice-snout is generally formed of a series of vertical pinnacles and sheets of ice, oriented transversely to the direction of ice movement. Each ice-cliff is separated from its preceding one by crevasses which narrow downwards into the glacier. The summits of the cliffs lean forward as they approach the glacier terminus and eventually crash down on to the terminal moraine, on to lower platforms of ice, or even on to the proximal slopes of the valley-train (figure 3.3, page 118). Little debris is contributed by these ice forms, except from lateral zones, for most of the debris falls into crevasses. The Bossons glacier in any case appears to carry relatively little surface

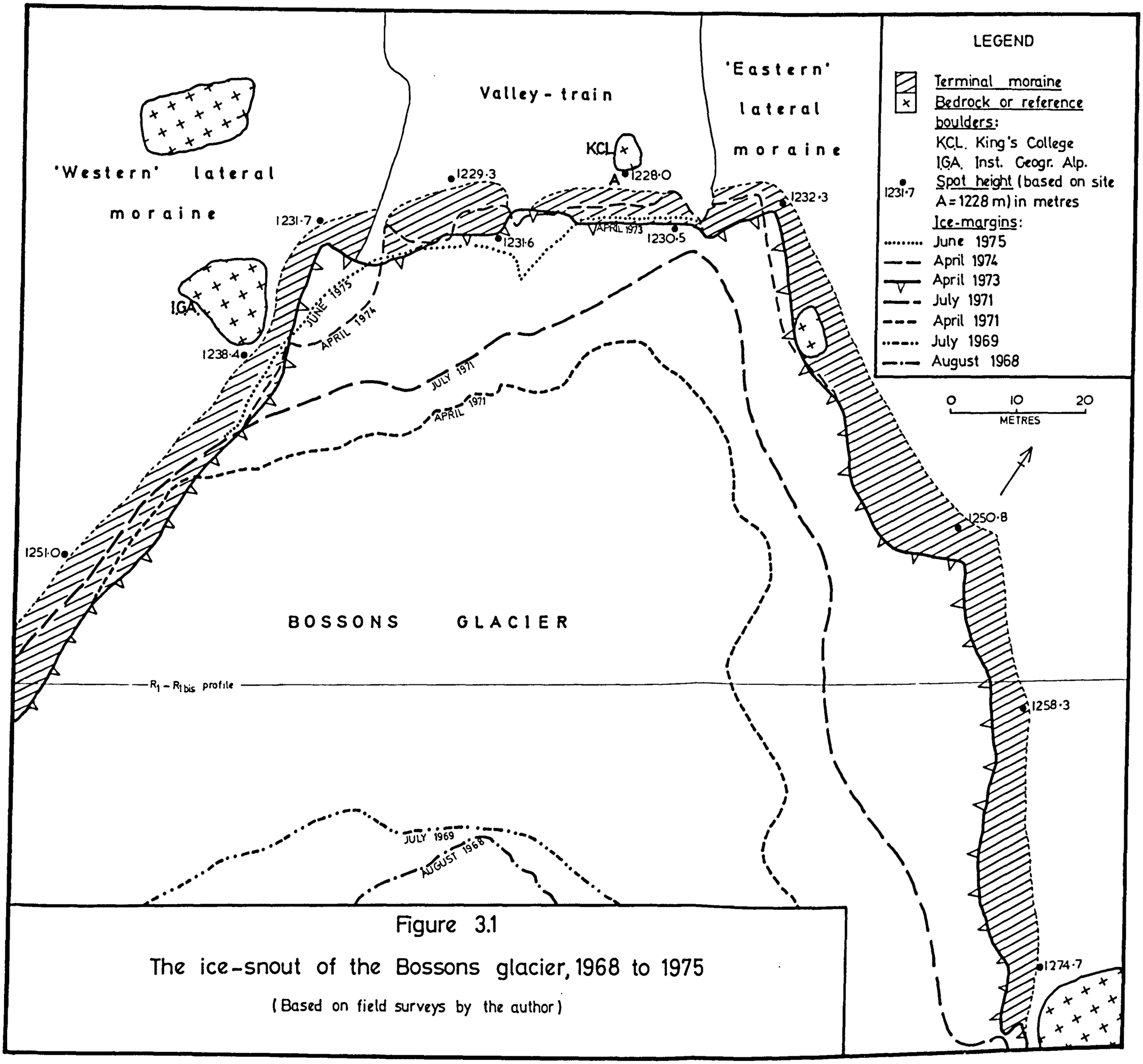


Figure 3.1

The ice-snout of the Bossons glacier, 1968 to 1975

(Based on field surveys by the author)

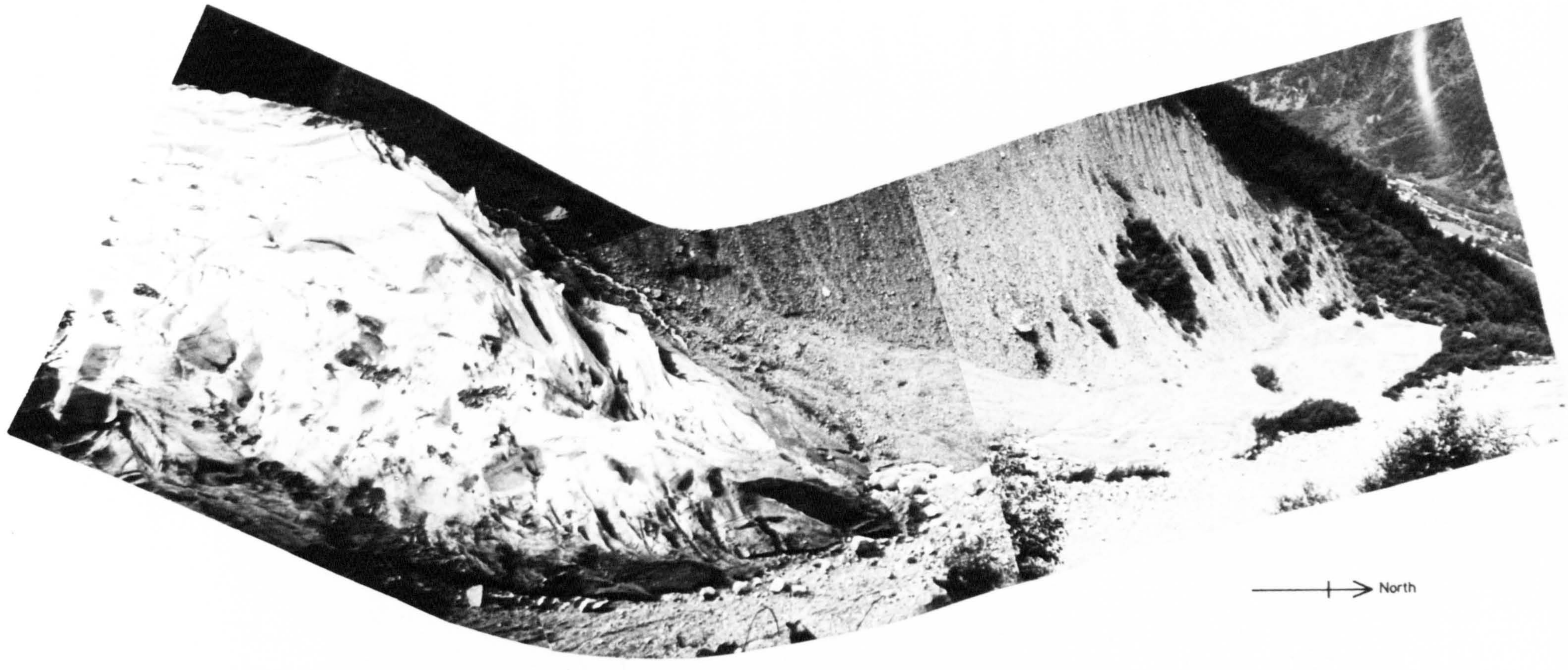


Figure 3.2 The ice-snout and valley-train of the Bossons glacier

debris and Veyret (1972) describes it as a "typical white glacier".

(2) Ice-margin 'platforms'. Ice-margin 'platforms' (figure 3.4, page 118) form at the ice terminus as a result of consistent ablation beneath a thin mantle of rock debris. These 'platforms' of ice are backed by ice-cliffs, and in the downvalley direction they tend to rise up and override the terminal moraine. The intervening depression between ice cliff and moraine (2, in figure 3.5, page 119) is then occupied by supraglacial meltwaters which accentuate the trough-like shape. The 'ice-platform' frequently overhangs the terminal moraine, so that the ice eventually cracks and slips down the distal slope of the moraine (figure 3.4, page 118). These detached sheets of ice are buried beneath morainic debris until they melt out or until meltwaters from the upper 'ice-platform' expose them.

Ice-cliffs are present at all times near the ice-margin, but 'ice-platforms' have formed in front of these steep ice features during periods of slow advance and particularly during the height of the meltseason. 'Ice-platforms' are generally confined to the western ice-margin where a more prominent terminal moraine causes the ice to rise up over the moraine ridge. Rising ice prevents the ice pinnacles from reaching the ice-margin intact.

3.1.2. ICE-MARGINAL DEPOSITS: MORAINES

The frequent advance of marginal ice over, and even into, the terminal and lateral moraines masks any distinctive divisions between 'ice' and 'moraine'. Debris-mantled ice and ice-cored moraine tend to merge at the ice-margins (see figure 3.5, page 119). The distal limits of the moraines can, however, be recognized by a significant break of slope.

The debris forming the terminal moraine is derived from three sources:

(1) Material lying beneath and in front of the glacier is shunted along and piled up into a heap by the advancing ice to form a 'push moraine' (Veyret, 1972). The moraine is therefore mainly composed of debris

Figure 3.3

Ice-cliffs at the Bossons glacier snout

Figure 3.4

Over-hanging ice-platform at the Bossons glacier snout

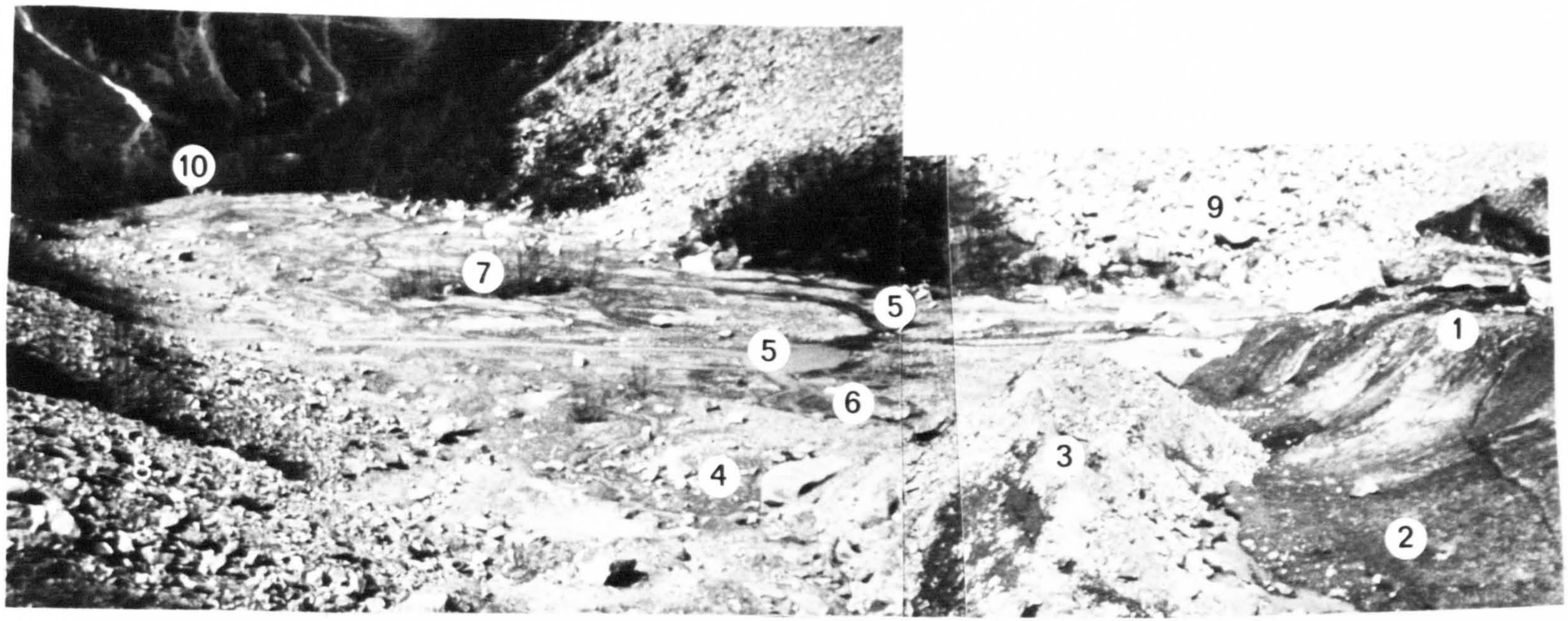


Figure 3.5 The south-west sector of the Bossons proglacial valley

1 Bossons glacier ice-margin
2 Ice-marginal depression
3 Terminal moraine
4 Débris fan
5 Kettle-hole
6 Infilled kettle-hole
7 Former terminal moraine
8 Western lateral moraine
9 Eastern lateral moraine
10 Outlet of valley-train

derived from basal moraine and older outwash deposits as they are over-ridden by ice.

(2) Englacial debris melts directly out of the ice and may be added to the moraine by meltwaters flowing over the ice-margin on to the moraine or, if the ice penetrates or overhangs the moraine, by debris falling straight down on to the moraine surface.

(3) Supraglacial debris is either washed off the ice-margin by meltwaters or is saturated to form mudflows which sludge on to the terminal moraine, forming a 'flow till' (e.g. Boulton, 1972).

The volume of debris which is built up to form the terminal moraine is surprisingly small. The moraines do not exceed 10m in height, and although usually much less than this, their height and extent has increased during the recent ice advance (see also Veyret, 1972). The lack of high moraines reflects both the general lack of debris carried within the glacier itself, and the removal of moraine deposits by meltwaters (see below). The moraine ridges are larger and more consistent on the west of the ice-margin where large amounts of debris are derived from the adjacent rock walls of the Montagne de la Côte, and where powerful meltwater streams are absent. There is far less supraglacial debris present on the east where the ice-margin is perforated by numerous meltwater stream outlets.

The terminal moraines, once deposited, may be modified by a variety of mechanisms:

(1) Meltwater stream activity can destroy the structure of the moraine by removing the fine sediments and leaving behind a coarse 'lag' deposit. Slumping and downward 'resettling' then occurs. Stream activity is usually localized, and eventually dissects the terminal moraine into a series of separate ridges (see also Galon, 1973a; Klimek, 1973).

(2) The collapse and subsequent melting of overhanging or penetrating ice shelves removes fine fractions in suspension down the distal moraine slope. Again structural

rearrangement takes place within the moraine as fines seep over and through the coarse structure and emerge at the base of the moraine (see also Kozarski and Szuprycynski, 1973).

(3) Meltwaters flowing over and through the moraine undergo diurnal freezing and thawing during the early and late stages of the meltseason. Differential movement of rock fragments and the reorientation of debris particles results from plastic deformation of the moraine deposits during the growth of veinlets, pockets and sheets of ice (Sigafoos and Hopkins, 1952; Pissart, 1969, 1970). Additionally, chunks of ice frozen to underlying and adjacent moraine debris remove the debris as they slip, or are washed down, the steep distal slope of the moraine. This mechanism is similar to that produced by 'anchor ice' in streams (Jenness, 1952).

(4) Changes in the position of the ice-margin can result in continued construction of the terminal moraine during an ice advance, or in its abandonment during an ice retreat. During an ice advance the original structure of the moraine is destroyed as it is driven downvalley, new debris is added, and a new, composite moraine is formed. Should the moraine be abandoned by ice melt, 'settling' of sediments occurs, accompanied by slumping and sediment creep, especially around melting ice and on the 'abandoned', oversteepened, proximal slope.

(5) Moraines which were 'abandoned' over twenty-five years ago have also been subject to post-depositional modification. These older moraines have been mentioned in section 2.6 in connection with the development of soil and vegetation. The high lateral moraines have been graded to form scree slopes, the mid-valley terminal moraine (vegetation habitat C,7 in figure 3.5, page 119) is being buried by aggrading outwash sediments, while all the older moraines have undergone some soil development and have been colonized by vegetation.

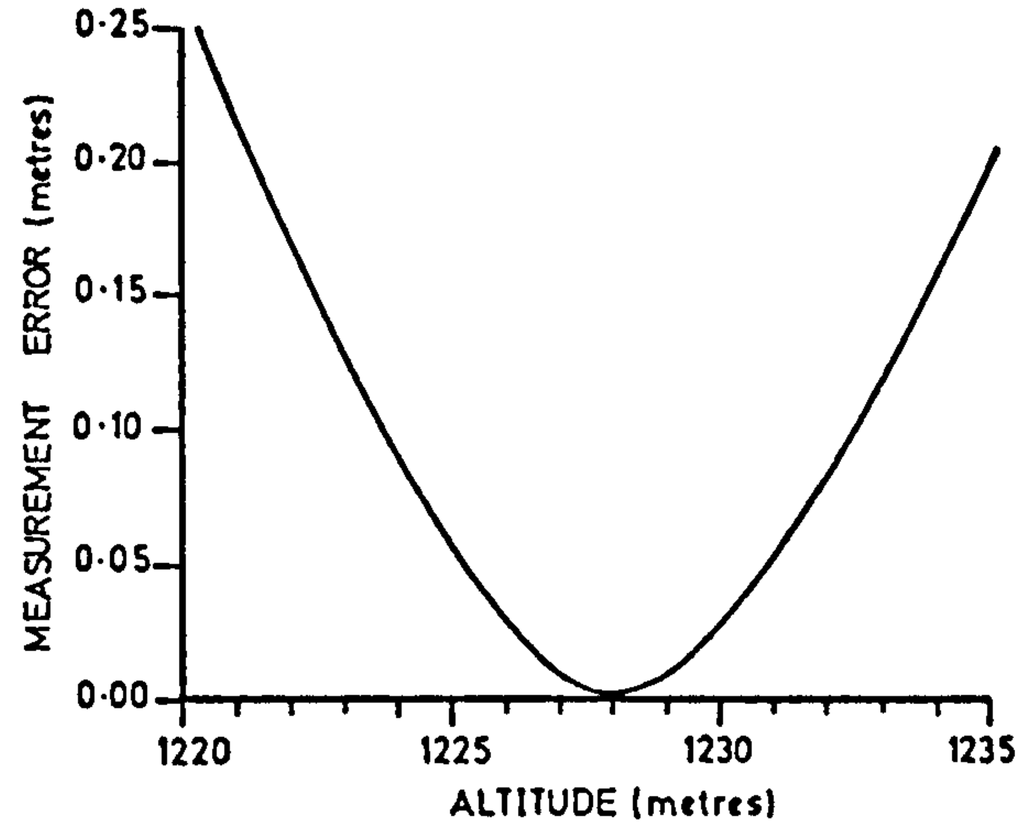
3.2. THE VALLEY-TRAIN

3.2.1. SURVEYING METHODS

Base-maps were constructed by field survey, since the regional topographic maps and aerial photographs lacked both detail and contemporaneity. The base-map was initially constructed by plane-table surveying, using alidade and tape, and produced at scales of 1:600 and later of 1:300. There are no spot heights or fixed points in the valley to which measured heights can be related, so an assumed altitude at site A (figure 3.7, page 124) was initially taken to be 1228m, based on estimates from larger-scale maps and from information given by Vivian (1969b) and Veyret (1972). The stream exit from the valley was subsequently considered as a more reliable fixed altitude because of the obvious proximal aggradation that has been taking place near site A. The distal stream exit (site S on figure 3.7) by contrast, appears to have remained at a constant level, taken to be 1221m. Earlier heights measured within the valley during plane-table surveying were too infrequent to construct anything other than general form lines. A more extensive network of over 200 heights was therefore constructed in 1973 and 1974 by taking Abney Level readings on two survey lines down the valley-train and at 5m intervals across the valley-floor (figure 3.7, page 124). Surveying started at site A which is close to the ice-margin, so that errors increase downvalley and upvalley from an altitude of 1228m (in 1973) (figure 3.6) (see also Konysz, 1973).

The positions of meltwater channels were also mapped. Diurnal and seasonal variations in stream width did not affect the mapping of channel limits which were plotted as the change in slope at the 'bankfull' level. During stages of higher water level, the margin of waterflow across a shoal surface was mapped (see plates 1a and b, 2,3; and Maizels, 1970).

Figure 3.6
Altitudinal error in
Abney Level measurements, April 1973



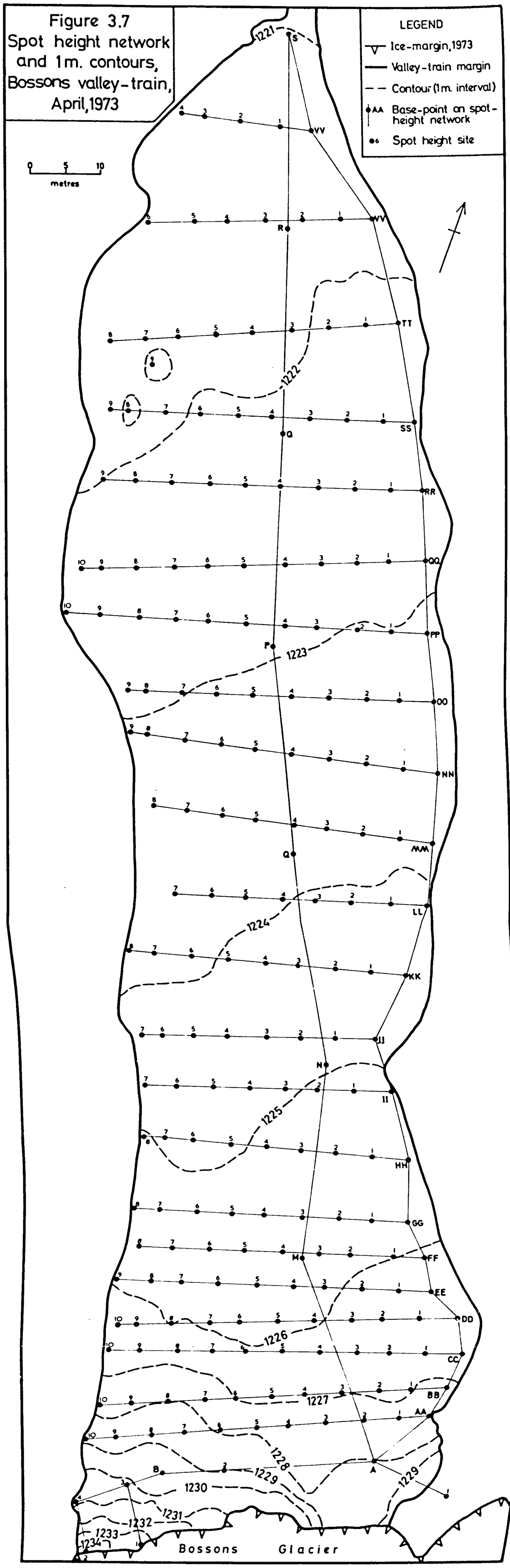
3.2.2. SURFACE MORPHOLOGY

3.2.2.1. Major features of surface morphology

The surface morphology of the proglacial valley-train exhibits a series of important features:

(1) Slopes are persistently steeper in areas closest to the ice-margin, where the terminal moraines and debris fans have accumulated to the greatest height (see plates 2 and 3; figure 3.8, page 125). The major debris fan lies in front of the western side of the ice-margin (4, in figure 3.5, page 119) and is composed of debris largely derived from material washed down from the more developed terminal moraines and steeper valley-side slopes situated on the west. Fan development is most rapid at the start of the meltseason, when debris of all sizes slides downslope, frequently as mudflows. During the height of the meltseason the fine fractions are continually washed out and the terminus of the fan shifts upslope as meltwater streams undermine and erode the foremost fan deposits. All distal edges of the fan are eroded during the meltseason, resulting in the formation of a channel confined between the debris fan and the foot

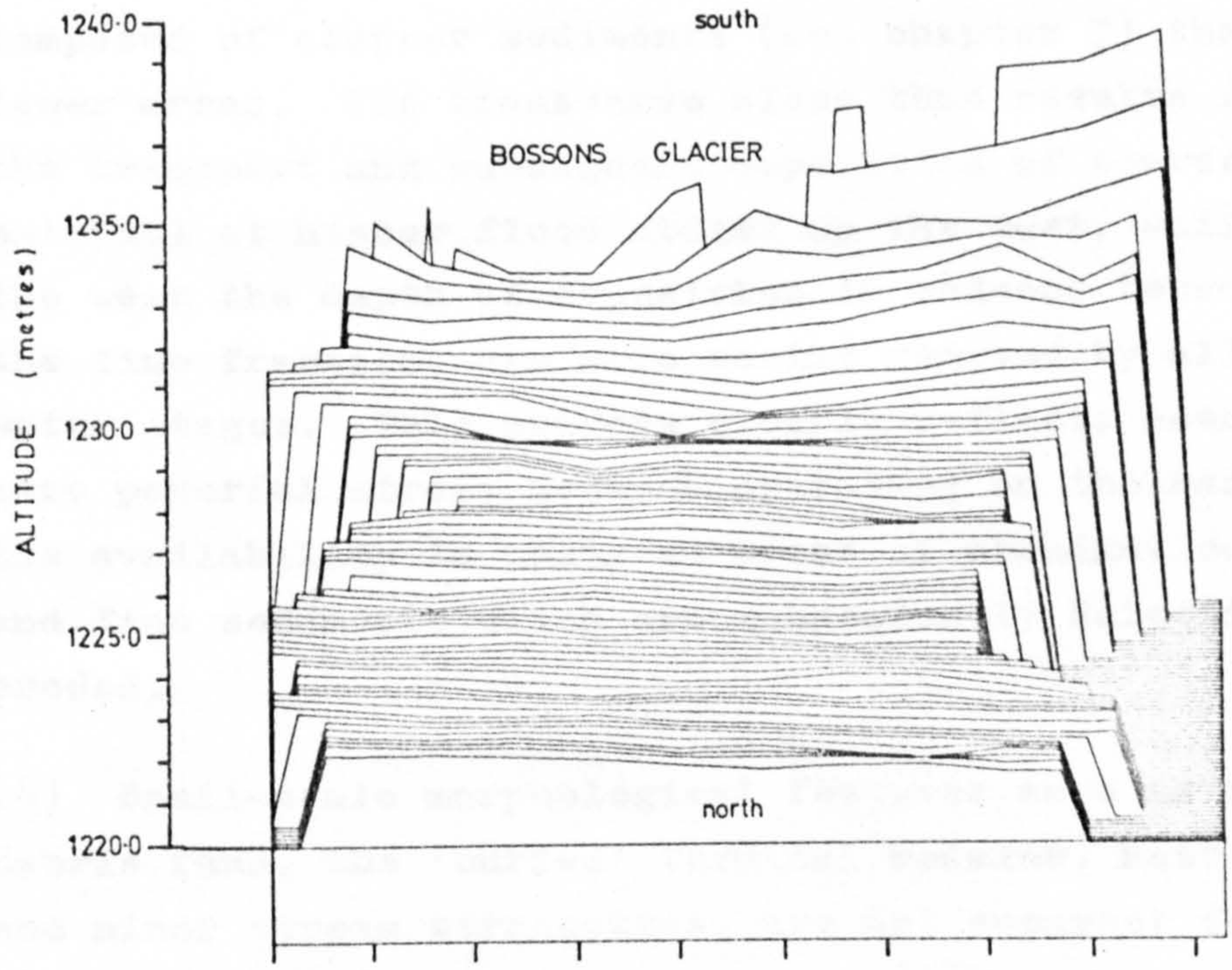
Figure 3.7
Spot height network and 1m. contours, Bossons valley-train, April, 1973



of the western lateral moraine. This channel is the major outlet which serves the marginal drainage system of the western ice-snout, north of the terminal moraine ridge.

Other channels tend to be well entrenched into the outwash deposits, and are often deep, narrow and steeply graded.

Figure 3.8
Topographic surface of the Bossons valley-train



Each division represents 5·0 m interval

(2) The gradient of the valley-train levels out downvalley and slopes of less than 2^{o} are frequently recorded over the remaining 150m of the valley-train (see section 3.3.). The deposits in this section of the valley-train are almost wholly of fluvioglacial origin, and include little debris which has fallen or slipped directly from the ice-margin and moraines. Debris has fallen from the neighbouring scree slopes in some lateral localities, however, to form minor debris fans. The surface of the valley-train is characterized by a complex network of anastomosing meltwater channels separated by banks or shoals of outwash deposits (see plates 1a and b, 2,3, figure 3.5, and Chapter 5).

(3) The pattern of contours reveals that the eastern half of the distal slopes of the valley-train generally lies at a higher level than the western half (see figure 3.7, page 124). The higher areas are also composed of coarser sediments (see chapter 7) than the lower areas. The transverse slope thus results from the transport and subsequent deposition of coarser material at higher flood stages on the east, while on the west the depth of deposition is reduced because the fine fractions are more easily removed by all melt-water stages. This process clearly reflects both the more powerful stream network operating on the east, and the availability in upstream areas of abundant coarse and fine sediments which are subsequently selectively eroded.

(4) Small-scale morphological features such as small debris fans, the 'buried' terminal moraine, kettle holes, and minor stream terracettes, are not apparent from the contour pattern, but they are nevertheless important proglacial features (see figure 3.5 and sections 3.5 and 3.6).

3.2.2.2. Changes in surface altitude from year to year

The variations in local topography from year to year have been ascertained using two methods. Firstly, the difference in altitude at each spot height site between 1973 and 1974 has been calculated, assuming a fixed altitude of 1221m at site S, the most downstream locality on the valley-train (figure 3.9, page 128). The terminal moraines have increased in height as the ice has advanced downvalley, and proglacial fluvioglacial and fan aggradation has been fairly extensive near the ice-margin. Some local erosion in downstream areas has occurred, especially within the formerly higher eastern shoal deposits. Changes in altitude have been greater than the estimated measurement error (figure 3.6, page 123) in almost all areas of the valley-train. Price (1971) also used a difference-of-heights method to determine the amount of change in altitude of proglacial outwash deposits of an Icelandic sandur. He attributed major reductions in altitude to the melting of buried ice and the formation of surface depressions whereas altitudinal changes in the surface in the Bossons valley are predominantly related to fluvial activity.

The second method compares the coefficients describing trend surfaces* applied to the height data in 1973 and 1974 (figure 3.10, page 129). The trend surfaces generalize the overall relief of the valley-train and differences in the coefficients indicate whether the pattern and degree of relief has significantly altered between April 1973 and April 1974. The initial coefficients for 1974 are up to 5.2m larger than for 1973 (table 3.1, page 130), indicating that aggradation and overall steepening of the pro-glacial valley-train has occurred in association with the slight glacier advance.

Some workers (e.g. see Kirby, 1969; Tarrant, 1970) have questioned the validity of employing the best-fit surface to altitude data. Most dissent, however, is directed against fitting trend surfaces to erosion platforms,

*Program compiled by I. and M. Clark.
See Appendix 5 for further discussion.

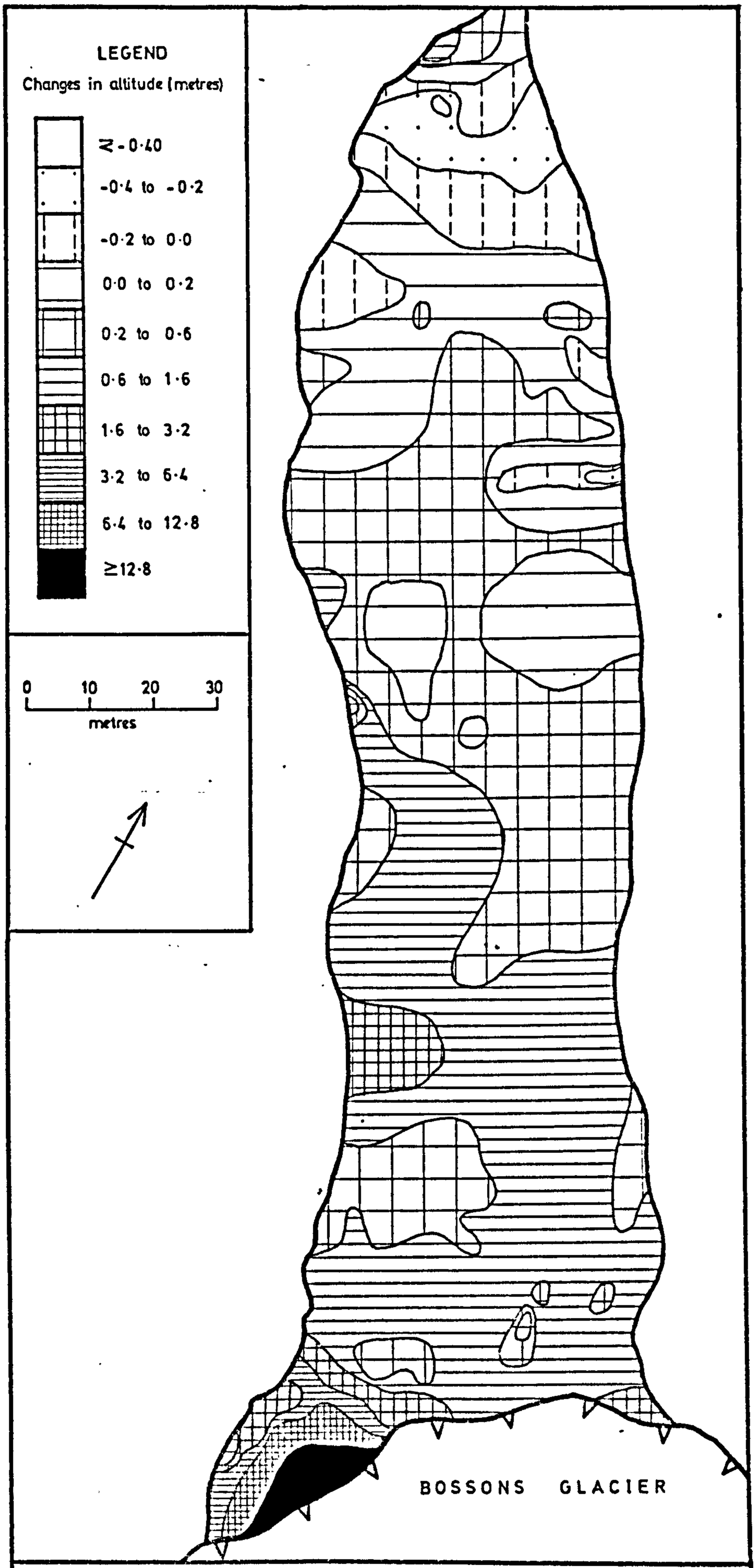
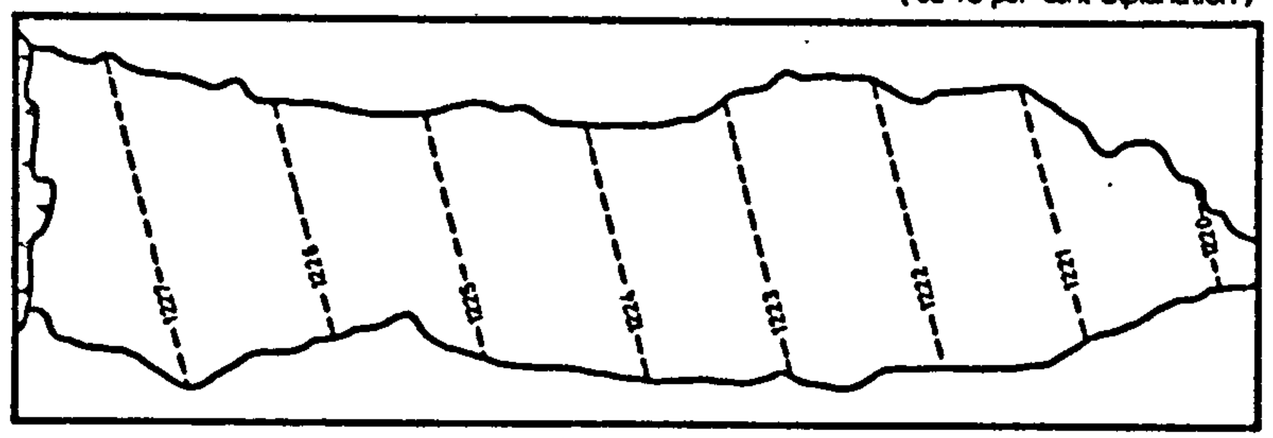


Figure 3.9 Changes in altitude on the Bossons valley-train between 1973 and 1974

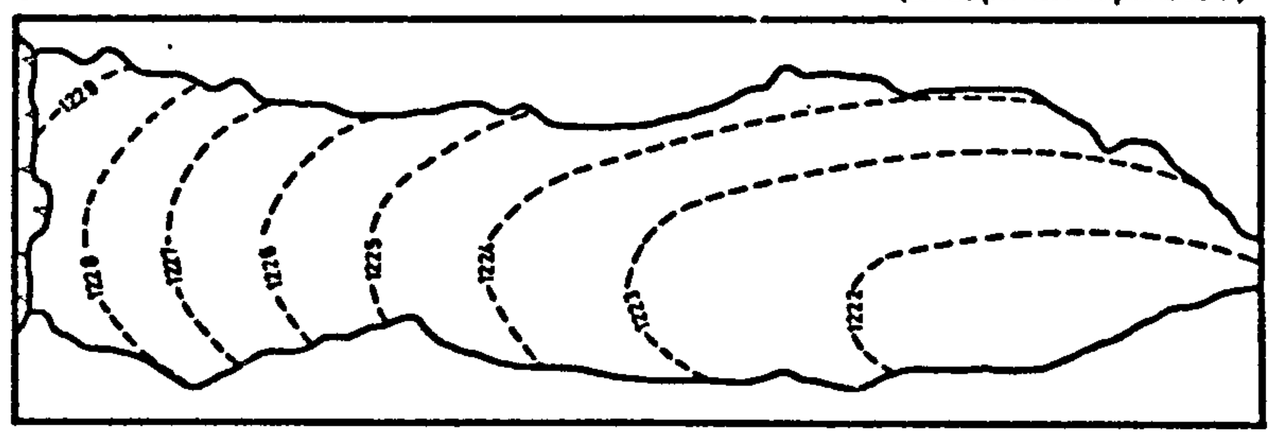
FIGURE 3.10 TREND SURFACES OF THE ALTITUDE OF THE BOSSONS VALLEY-TRAIN, 1973 AND 1974

1973

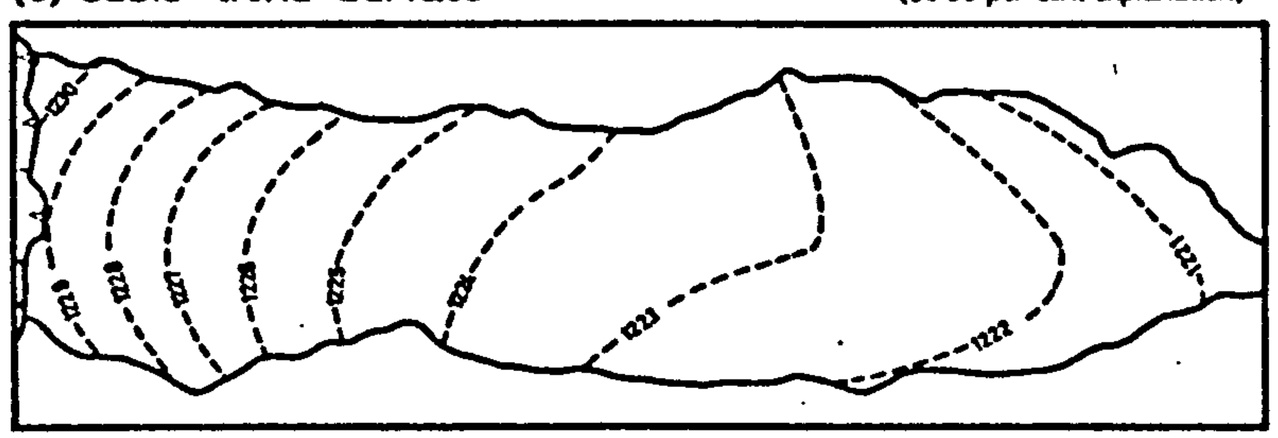
(a) Linear trend surface (82·78 per cent explanation)



(b) Quadratic trend surface (92·20 per cent explanation)

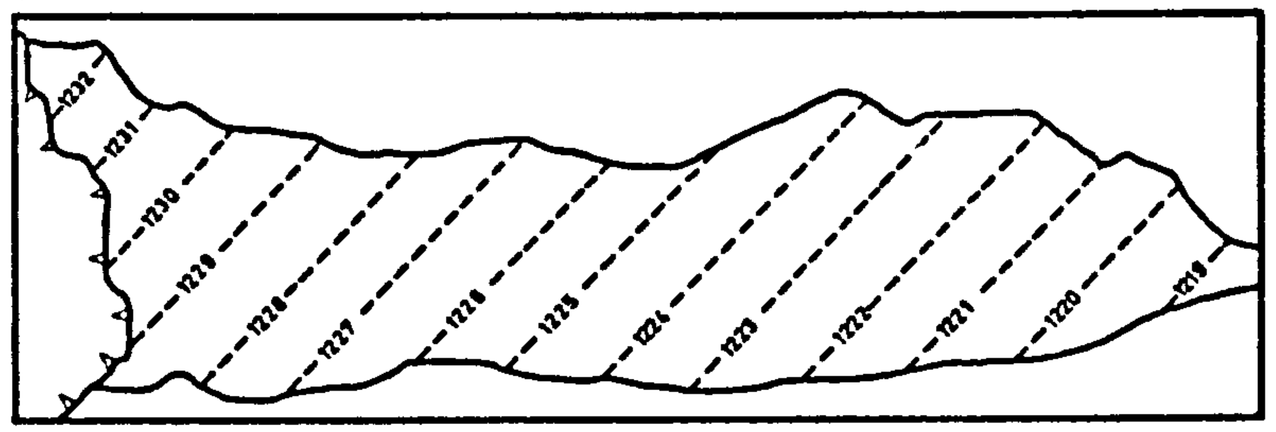


(c) Cubic trend surface (95·95 per cent explanation)

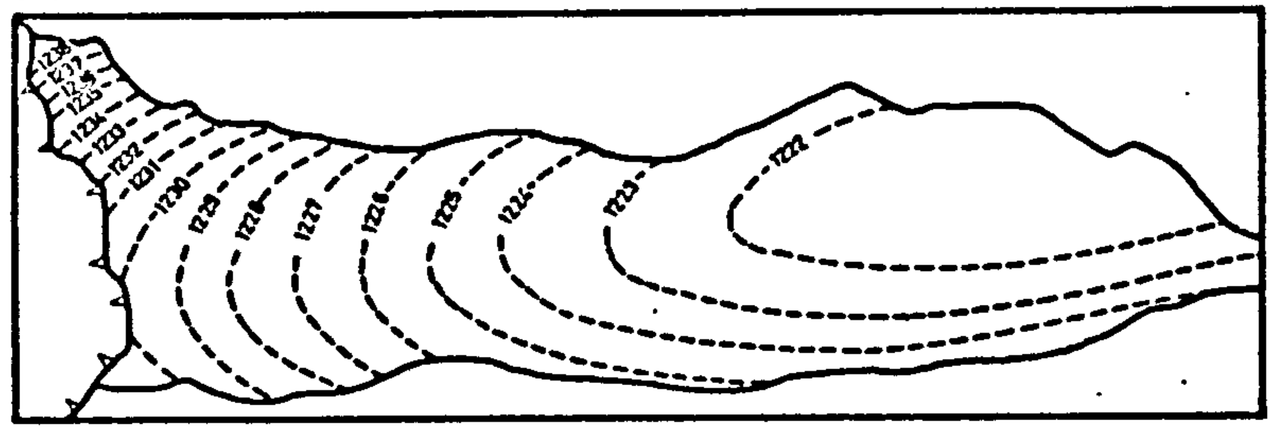


1974

(a) Linear trend surface (57 04 per cent explanation)



(b) Quadratic trend surface (71·83 per cent explanation)



(c) Cubic trend surface (78 81 per cent explanation)

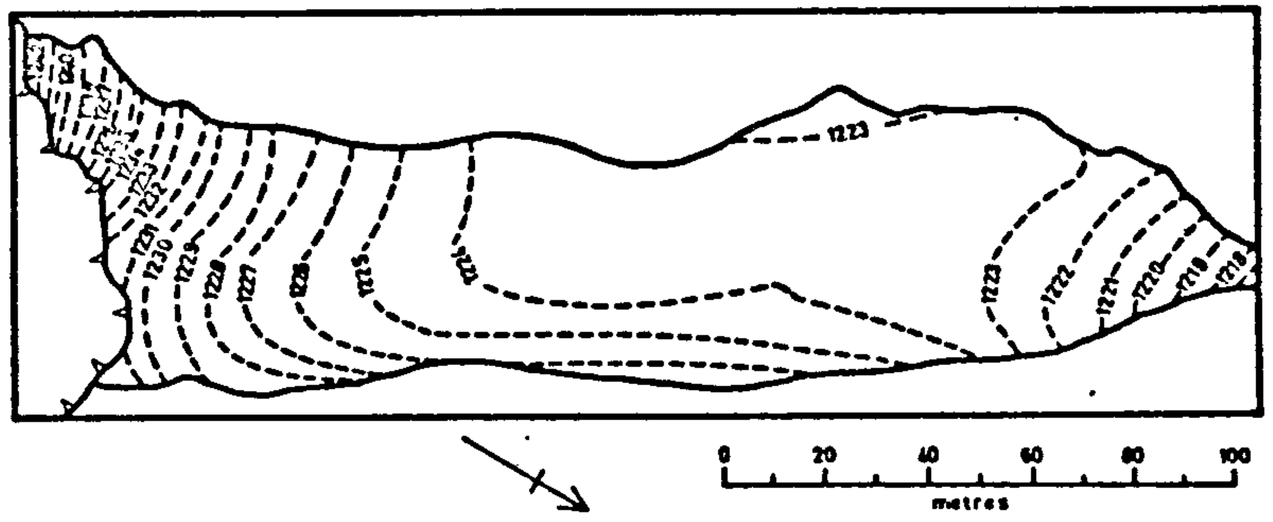


Table 3.1

Trend surface analysis of the altitude of the Bossons valley-train, 1973 and 1974

1. Analysis of variance tables

	1973					1974				
Source	Sum of squares	Degrees of freedom	Mean square	F-ratio	signif. level	Sum of squares	Degrees of freedom	Mean square	F-ratio	signif. level
Linear	1004.90	2	502.43	519.18	0.001	1863.00	2	931.48	97.58	0.001
Residual	209.03	216	0.97			1403.20	147	9.55		
Quadratic	1119.20	5	223.83	503.32	0.001	2346.00	5	469.20	73.43	0.001
Difference	114.31	3	38.10	85.68	0.001	483.02	3	161.01	25.20	0.001
Residual	94.73	213	0.44			920.18	144	6.39		
Cubic	1164.70	9	129.41	550.01	0.001	2574.00	9	286.00	57.84	0.001
Difference	45.55	4	113.87	48.40	0.001	227.98	4	57.00	11.53	0.001
Residual	49.18	209	0.24			692.19	140	4.94		
Total	1213.90	219				3266.20	150			

2. Percentage of total sum of squares

Component	1973	1974
Linear	82.78	57.04
Quadratic	92.20	71.83
Cubic	95.95	78.81

3. Solution of equations

Surface	1973	1974
Linear	$z_L = 1227.40 - 0.81x + 0.167$	$z_L = 1229.30 - 0.16X + 0.15Y$
Quadratic	$z_Q = 1230.80 - 1.36x - 2.17Y + 0.10X^2 - 0.13XY + 0.54Y^2$	$z_Q = 1233.50 - 0.29X - 0.38Y - 0.01XY + 0.03Y^2$
Cubic	$z_C = 1232.20 - 2.59x - 2.86Y + 0.31X^2 + 0.27XY + 0.78Y^2 - 0.02X^3 - 0.01Y^3 + 0.05X^2Y - 0.16XY^2$	$z_C = 1237.40 - 0.61X - 1.07Y + 0.02X^2 - 0.03XY + 0.12Y^2$

variable X = north-south component, variable Y = east-west component, variable Z = altitude component, to two significant decimal figures (other coefficients omitted)

for the fitted surface represents not the concordant summit level, but the average of all heights on the ground. The latter is a more useful measure of the Bossons valley-train surface, as it allows the recognition of areas of ground which are higher or lower 'than expected' by the plotting of positive and negative residuals. The residual maps for 1973 and 1974 (figure 3.11, page 132) indicate areas lying above and below the average surface, and these are located in similar areas in both 1973 and 1974. Positive or higher ground lies close to the ice-margin, representing the high depositional surfaces of the terminal moraine and debris fan. The residuals are greater in 1974 than in 1973, for proximal aggradation was more pronounced by 1974. Thus while aggradation has been continuing it has also remained concentrated in similar areas. The sources of added debris have therefore also remained constant.

3.3. THE LONGITUDINAL PROFILE OF THE VALLEY-TRAIN

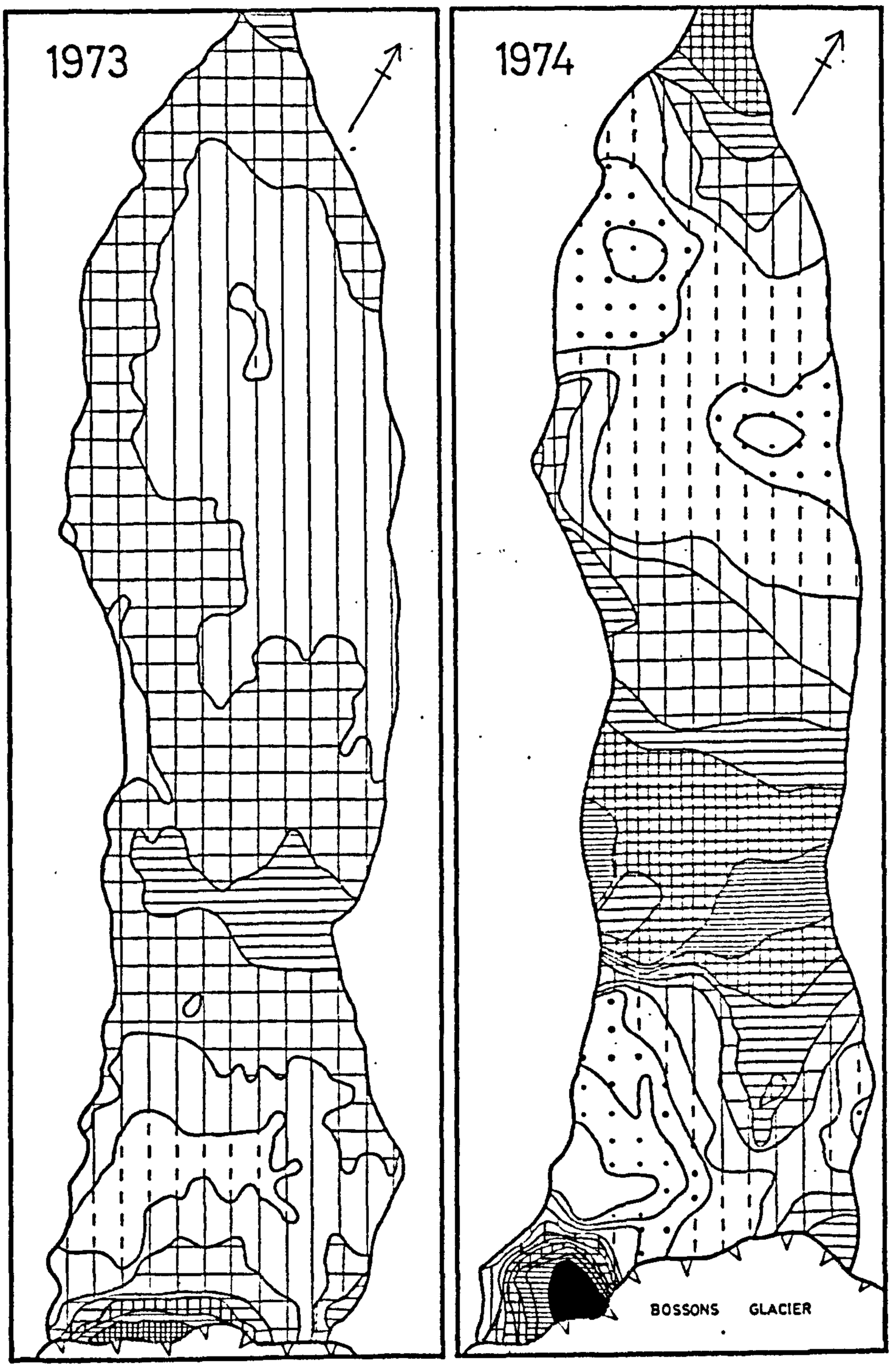
3.3.1. THE SHAPE OF THE LONGITUDINAL PROFILE

Comparison of valley-train gradients and identification of changes in slope are facilitated by the construction of longitudinal profiles. The long profiles were located along the median line of the valley floor. The longitudinal profile of the valley-train is of concave-upward form, being steep near the ice-margin and levelling out downvalley (figure 3.12, page 133). Altitudinal data have been collected from six field surveys completed between 1968 and 1974 (figure 3.13, page 134).

The double-log equations which have been fitted to profiles of the Bossons valley-train (table 3.2, page 136 and figure 3.14, page 135), each represent a highly significant correlation between elevation and distance from the ice-margin. However, a single equation does not fit the plotted data as closely as suggested by the correlation coefficients in table 3.2, for generally there are at least

FIGURE 3.11

ALTITUDE RESIDUALS OF THE CUBIC TREND SURFACES OF THE BOSSONS VALLEY-TRAIN, 1973 AND 1974



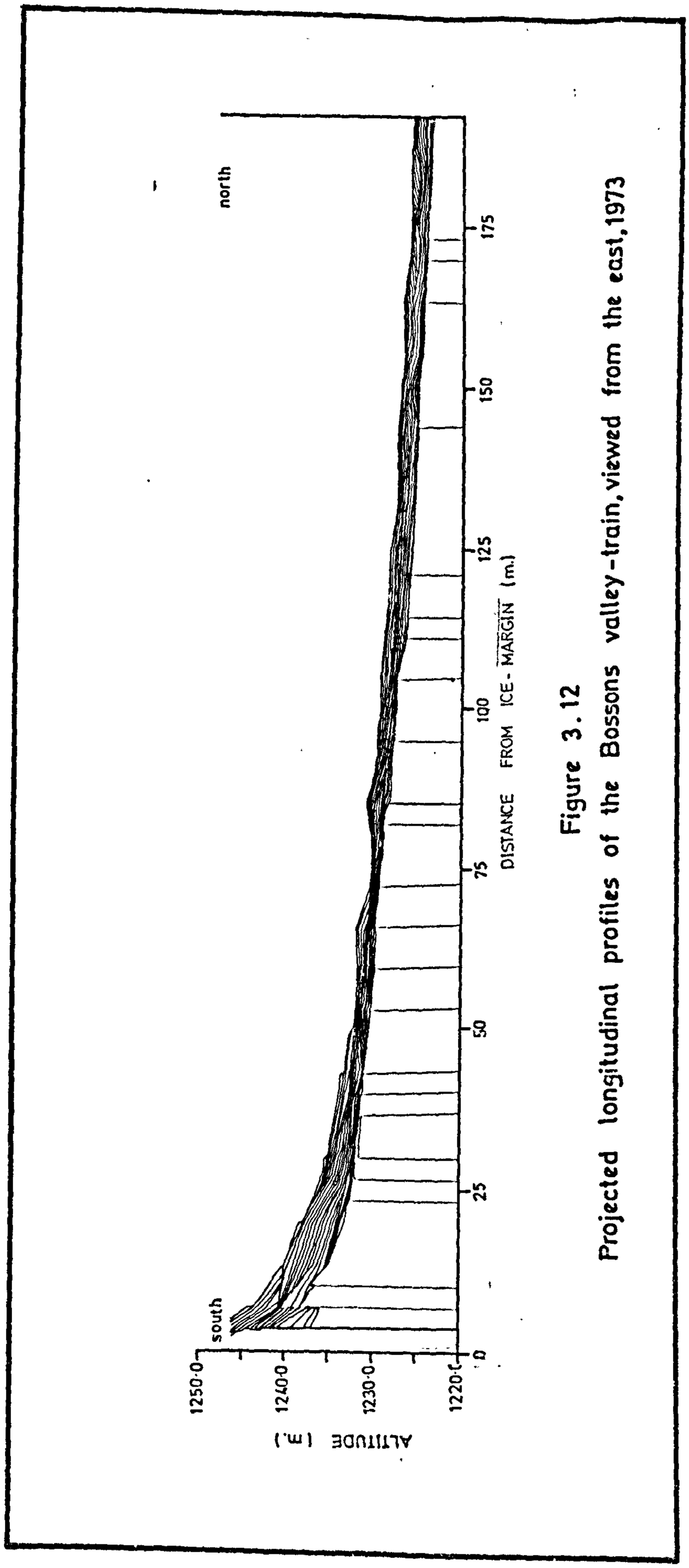


Figure 3.12

Projected longitudinal profiles of the Bossons valley-train, viewed from the east, 1973

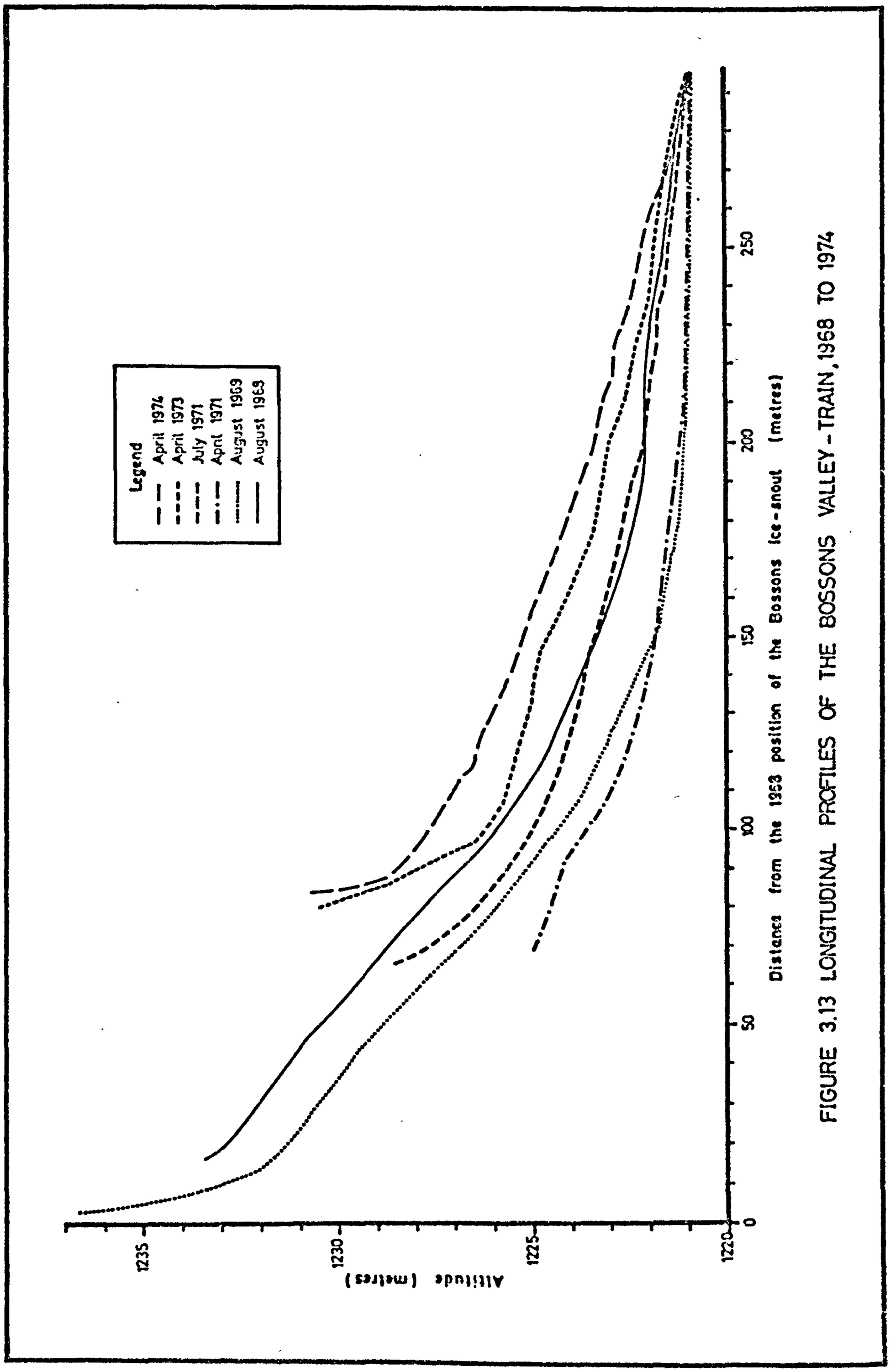


FIGURE 3.13 LONGITUDINAL PROFILES OF THE BOSSONS VALLEY-TRAIN, 1968 TO 1974

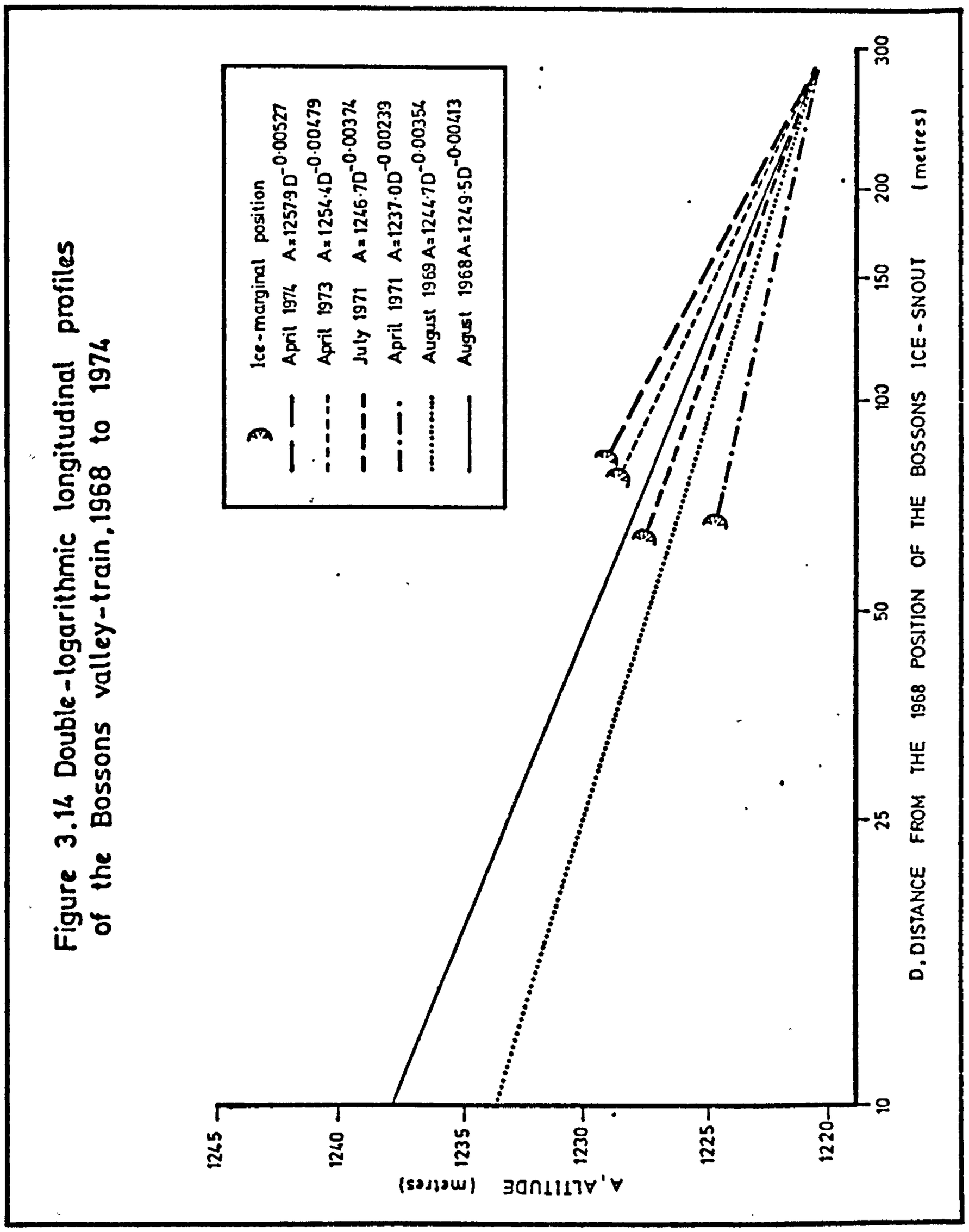


Figure 3.14 Double-logarithmic longitudinal profiles of the Bossons valley-train, 1968 to 1974

Table 3.2

Longitudinal profiles of the Bossons valley-train (log-log height-distance data)

Date of measurements*	r (1)	r^2 (2)	Signif. level (3)	a (4)	b (5)	$S.E._b$ (6)
August 1968	-0.988	0.976	0.001	1249.5	-0.00413	0.00013
August 1969	-0.979	0.958	0.001	1244.7	-0.00354	0.00014
April 1971	-0.961	0.885	0.001	1237.0	-0.00239	0.00015
July 1971	-0.985	0.858	0.001	1246.7	-0.00374	0.00014
April 1973	-0.973	0.813	0.001	1254.4	-0.00479	0.00025
April 1974	-0.960	0.922	0.001	1257.9	-0.00527	0.00034

where (1) r is the correlation coefficient;
(2) r^2 is the coefficient of determination;
(3) the significance level is based on tables of the distribution of F (Blalock, 1960);
(4) 'a' is the intercept value, converted to metres;
(5) 'b' is the slope value;
(6) $S.E._b$ is the standard error of b.

*The dates of measurement do not necessarily represent the date of formation of the profile, since the profiles are usually composite in origin, different areas undergoing erosion and aggradation at different times.

two well-defined, exponential slope segments.[1] Other workers have also recorded several slope segments on proglacial outwash profiles (e.g. Hoppe, 1959; Fahnestock, 1963,1969; McDonald and Banerjee, 1971; Boothroyd, 1972; Broscoe, 1972; Church, 1972), while many workers have observed similar slope

---

1. Authors standardize the form of equation representing an exponential profile by using a power function based on the gradient of the profile and described by:

$S = S_o e^{-cx}$ (derived from $\log_{10} S = \log_{10} S_o - 0.434\ cx$)

where S = Slope (dy/dx)
y = altitude
x = distance from source (or ice-margin)
S_o = slope at x = o
c = coefficient of proportionality, expressed as x, where $S = S_o/2$
and e = base of natural logarithms (2.718)

Values of 'c', the distance from the source at which the original gradient is halved, are useful indices for comparison of different longitudinal profiles. This is discussed in greater detail in chapter 7.

discontinuities on 'normal' (other than fluvioglacial) stream profiles (e.g. Krumbrein, 1937; Shulits, 1941; Yatsu, 1955; Hack, 1957).

The shape of the longitudinal profile of the valley-train surface is largely related to the balance between stream energy and the amount of stream load when the stream was flowing over that surface. An excess of debris beyond the amount that the stream is competent to remove will result in aggradation of sediment. Excess debris is continually added to the proximal slopes of the valley-train from the ice-margin, the moraines, the debris fans, and the heavily-laden meltwaters emerging from the glacier under hydrostatic pressure. A steep aggradational slope is built up directly in front of the ice-snout. The decline in slope downstream is partly related to a decrease in the size of particles forming the stream load, for a decrease in particle size allows the release of stream energy from the transport-ation of sediment to the use of energy in degradation of the stream bed. However, the gradient does not appear to decline regularly downstream, but is marked by several sudden breaks of slope. The breaks of slope represent a relatively large change in the rate of addition or of removal of debris (Yatsu, 1955; Culling, 1957a,b, 1960).

The major break of slope present on the Bossons valley-train occurs between (1) the proximal slope, which is largely composed of coarse debris transported short distances from the moraines and ice-margin before being dumped, and (2) the distal slope, where deposits are almost solely fluvioglacial in origin (see table 3.3). The junction between the two major slope segments thus represents a change in the sources of debris and the competence of streams to remove different proportions of the debris.

Table 3.3

Average gradient of slope segments on the Bossons valley-train, 1968 to 1974

Date of survey	'proximal' slope segment (m/m)	'distal' slope segment (m/m)
August 1968	- 0.0780	- 0.0193
August 1969	- 0.0858	- 0.0106
April 1971	- 0.0400	- 0.0064
July 1971	- 0.0708	- 0.0188
April 1973	- 0.1060	- 0.0255
April 1974	- 0.1057	- 0.0314

3.3.2. CHANGES IN THE LONGITUDINAL PROFILE, 1968 to 1974

The elevation of the valley-train has changed systematically over the seven-year period, although the shape of the profile has remained essentially constant. Changes of elevation have reflected changes in the proglacial pattern of aggradation and erosion during the recent period of glacier advance. Summary measures, presented in table 3.4, represent only the 'net result' after a period of undocumented, local erosional and depositional events (see also Stewart and Lamarche, 1967).

Table 3.4

Aggradation of valley-train surface and degree of glacier advance

Period of measurement	Average relative change in surface elevation (metres)	Glacier advance (metres)
Aug.1968 to Aug. 1969	-1.20	+3.0
Aug.1969 to April 1971	-0.20	+75.0
April 1971 to July 1971	+1.09	+ 7.0
July 1971 to April 1973	+0.95	+21.0
April 1973 to April 1974	+0.50	+ 6.0

Down-cutting of the valley-train occurred while the ice-margin remained almost stationary, between 1968 and 1969. Streams were not overladen with much material added from new sources or new areas covered by ice, and stream energy appears to have been concentrated into degrading most of the surface of the valley-train. Removal of debris and lowering of the surface is most likely to have occurred during periods of flood discharge. The maximum advance of 75 metres between August 1969 and April 1971 caused the glacier to override large quantities of proglacial outwash and morainic debris, and to bring higher meltwater discharges to the new ice-marginal localities. Extensive net proglacial aggradation occurred during the meltseason following this major advance, and has been slowly continuing during the more recent, but less extensive, advances.

Continued aggradation is also made evident by the progressive burial of the abandoned terminal moraine beneath outwash sediments. Pits were dug around the margins of the moraine in April 1974 revealing 'fresh' moss and soil to depths of 30cm. Vegetation and boulders on the terminal moraine are also being progressively submerged under flood deposits (figures 3.15a and b, page 140).

Localized erosion has also taken place, the chief result being the formation of numerous minor 'terracettes' cut into shoal surfaces. Terracettes may form in any part of the valley-train, but in 1973 and 1974 were concentrated in downstream localities. The differences in elevation between these two periods of measurement indicate that the 1974 profile was lower than the 1973 profile only in the 30 to 40m nearest the distal end of the valley-train (figure 3.16, page 141).

Maximum aggradation occurred in the proximal zone, i.e. closest to the sources of added debris. A similar pattern is found in all the series of profile measurements, namely, maximum changes in valley-train elevation occur on the proximal slopes, and the changes decrease steadily downstream. Regression analysis of changes in surface

Figure 3.15
Progressive burial of the former terminal moraine,
Bossons valley-train

(a) April 1971

(b) April 1974

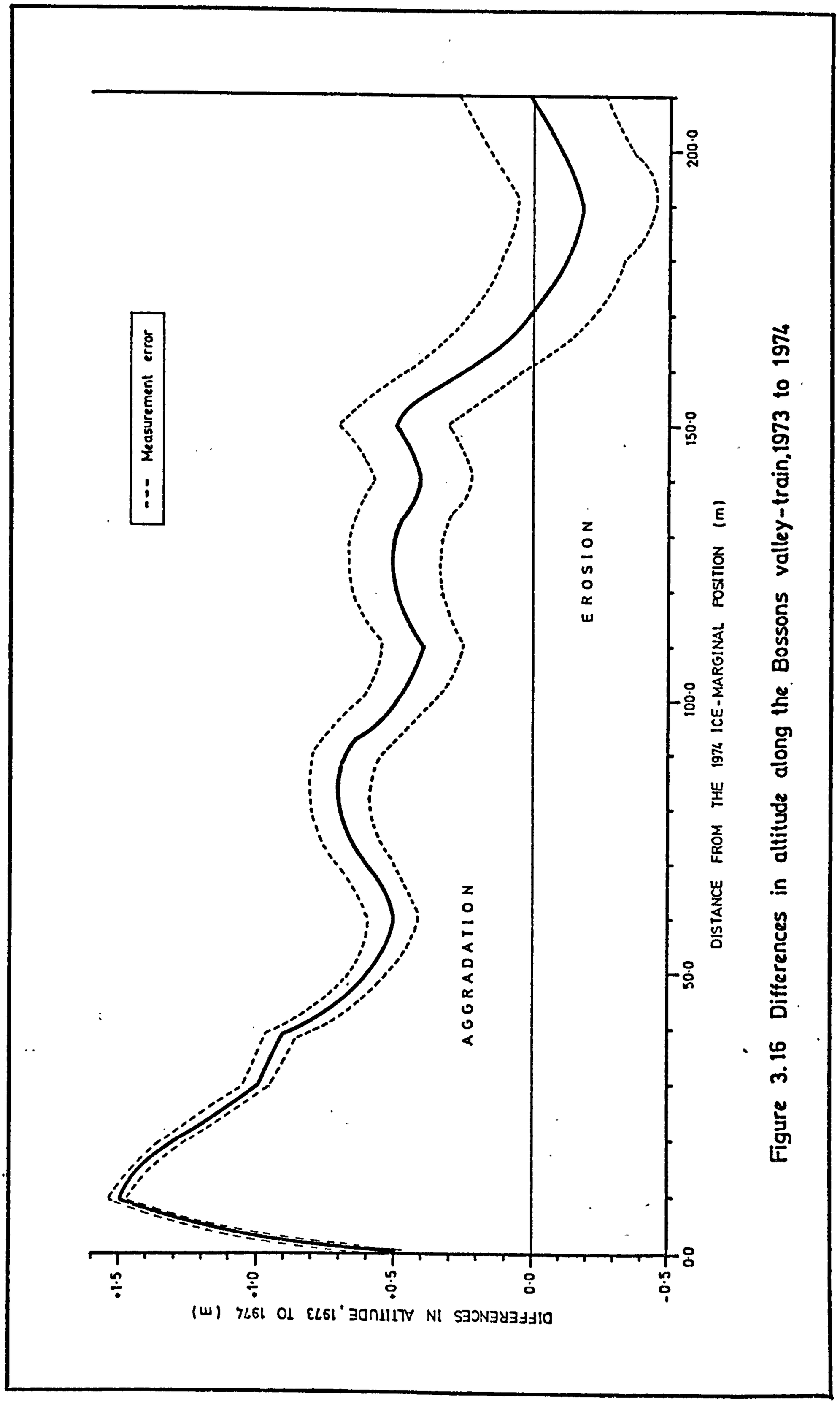


Figure 3.16 Differences in altitude along the Bossons valley-train, 1973 to 1974

elevation between successive measurements with increasing distance from the ice-margin indicates that the degree of change of elevation decreases significantly downvalley. This is partly accounted for by the fixed base-level at the distal edge of the valley-train, but is mostly the result of streams apparently grading towards the level of this distal outlet. It seems likely that the relative stability in elevation at the outlet reflects the presence of bedrock close to the surface, probably in the form of a rock bar buried beneath a thin mantle of outwash sediments (see also section 3.4). A further result of continued proximal aggradation has been the relative rise in elevation of the base of the ice-snout despite its advance into the formerly lower altitudes of the valley-train. The elevation of the ice-snout increased by about 5m between April 1971 and April 1974, although it advanced over 30m downvalley.

Steepening of the valley-train gradients has also accompanied the recent glacier advance. Most of the increase in slope has resulted from extensive aggradation in the proximal zone since April 1971 (see 'b' values in table 3.2, page 136 and table 3.3), but figures show that the distal slope of the valley-train has also progressively increased (table 3.3). An increase in the gradient of the proximal slope has thus been associated with an increase in the distal slope (figure 3.17, page 143), largely as a result of increased volumes of proglacial debris accumulating in a proglacial valley which is decreasing in area as the glacier advances over it.

The location of the 'transition zone' or 'intermediate zone' (Fahnestock, 1969) representing the junction between the two major slope segments, may be partly dependent on the gradient of the proximal slope (figure 3.18, page 143). The steeper proximal gradients tend to be associated with shorter proximal slope segments (table 3.3; see also Price, 1973) so that as the glacier has advanced and the snout has increased in elevation, the 'transition zone' has remained within a constant 30m-wide belt (table 3.5).

Figure 3.17
Gradients of proximal and distal slope segments on the Bossons proglacial valley-train, 1968 to 1974

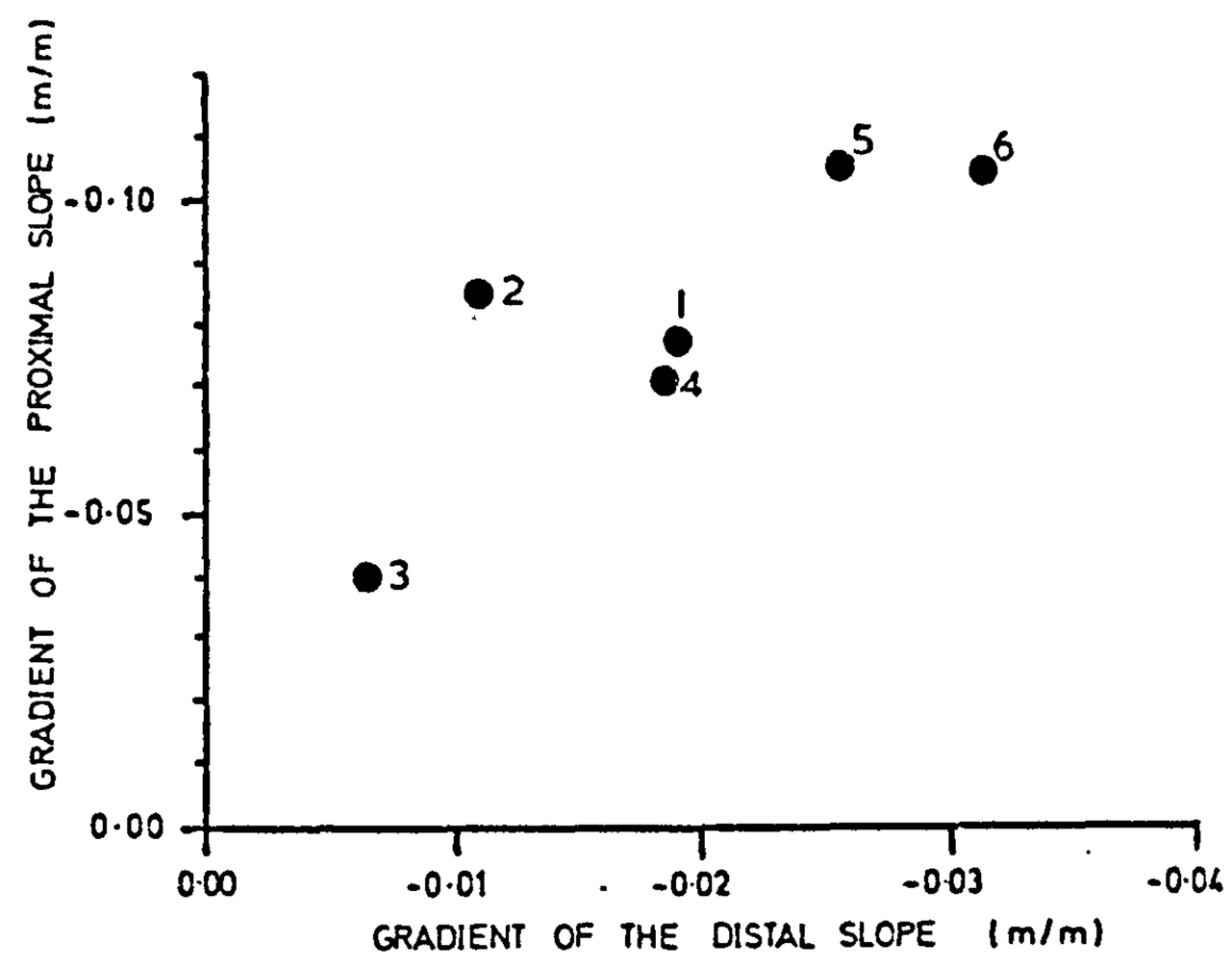
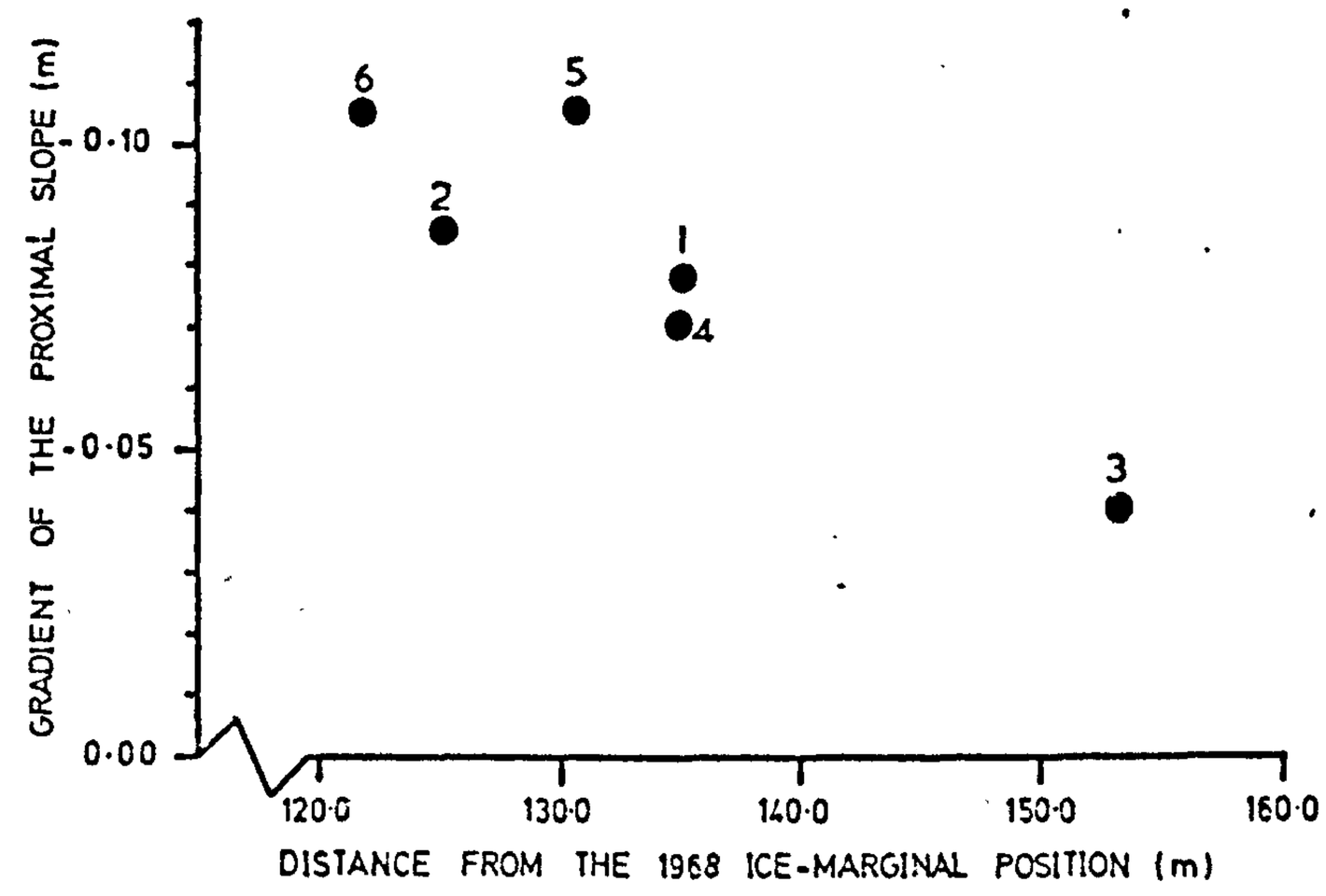


Figure 3.18
Location of the break of slope between the proximal and distal slope segments



1 August 1968
2 August 1969
3 April 1971
4 July 1969
5 April 1973
6 April 1974

Table 3.5

The average distance from the ice-margin of the major break in slope on the Bossons valley-train, 1968 to 1974

Period of measurement	Approximate distance of break of slope (metres) (a) from the ice-margin	(b) from the 1968 ice-marginal position
August 1968	135	135
August 1969	122	125
April 1971	75	153
July 1971	50	135
April 1973	25	131
April 1974	15	122

Historical evidence suggests that this zone may mark the downstream limit of a former debris fan formed downvalley of a terminal moraine deposited in the 1870's. The progressive addition of debris to the proximal slope increases its elevation by accumulating over the older deposits and increases the gradient by adding more material at the crest than can be removed. The slopes do not extend beyond the 'transition zone' because at that point meltwaters begin to undercut the foot of the slope. The break of slope is shifted slightly upvalley during the meltseason as streams cut back into the proximal deposits. This process can operate rapidly and effectively in the heterogeneous and unconsolidated materials which characterize the outwash deposits.

3.4. THE CROSS-PROFILE OF THE PROGLACIAL VALLEY

The cross-profile of the valley-train (figure 3.8, page 125) exhibits higher ground on the east, and lower ground bounding the western margins of the valley-train, except on and near the proximal slopes where debris fans have formed, streams have become entrenched and 'kettle-holes' have developed.

Construction of the cross-profile of the whole proglacial valley may provide a means of estimating the probable thickness of the outwash deposits. The method involves fitting parabolic curves to the valleyside slopes by trial and error and extrapolating them beneath the present surface. This method has frequently been applied in the study of glacially eroded valleys (e.g. Svensson, 1959; Graf, 1970, and personal communication 1975), but never to essentially depositional slopes (see also valley cross-sections discussed in Lewis, 1947 and Bezinge et al., 1973). The main drawback is clearly that the lateral moraine slopes have been modified by basal talus formation. The curves have therefore only been fitted to the upper slopes, although these have presumably been worn back slightly from their original ice-marginal position. The heights of the slopes were determined from field surveys.

The most closely fitting parabolic curves were $y = 1210 + 0.0049x^2$ and $y = 1220 + 0.0044x^2$ (figure 3.19, page 146) where y = altitude and x = distance from centre of the valley-train . The equations indicate that bedrock probably lies between 8 and 18m beneath the proximal surface and between 1 and 11m beneath the distal surface. A value of only 1m would support the proposal for the existence of a rock bar close to the outwash surface (see above). The rock base near the ice-snout therefore probably lies at an altitude of between 1220 and 1210m, with a maximum overall downvalley gradient of only 0.045. This gradient is lower than that of the aggraded surface of the valley-train (see table 3.3, page 138). A bed-rock level at 1210m is supported by the calculation of the valley-side profiles using a semi-logarithmic transformation (figure 3.20, page 147). The upper slopes of these profiles intersect beneath the outwash surface at an altitude of 1211m. The exponents of the double-logarithmic plots (Svensson, 1959) of the original valley-side slopes (figure 3.21, page 147) are also very similar, indicating that the valley is indeed almost symmetrical. However, the values of the exponents fall below the value 2 which describes a true parabola, and the fact that the exponents of the parabola based on a 1220m

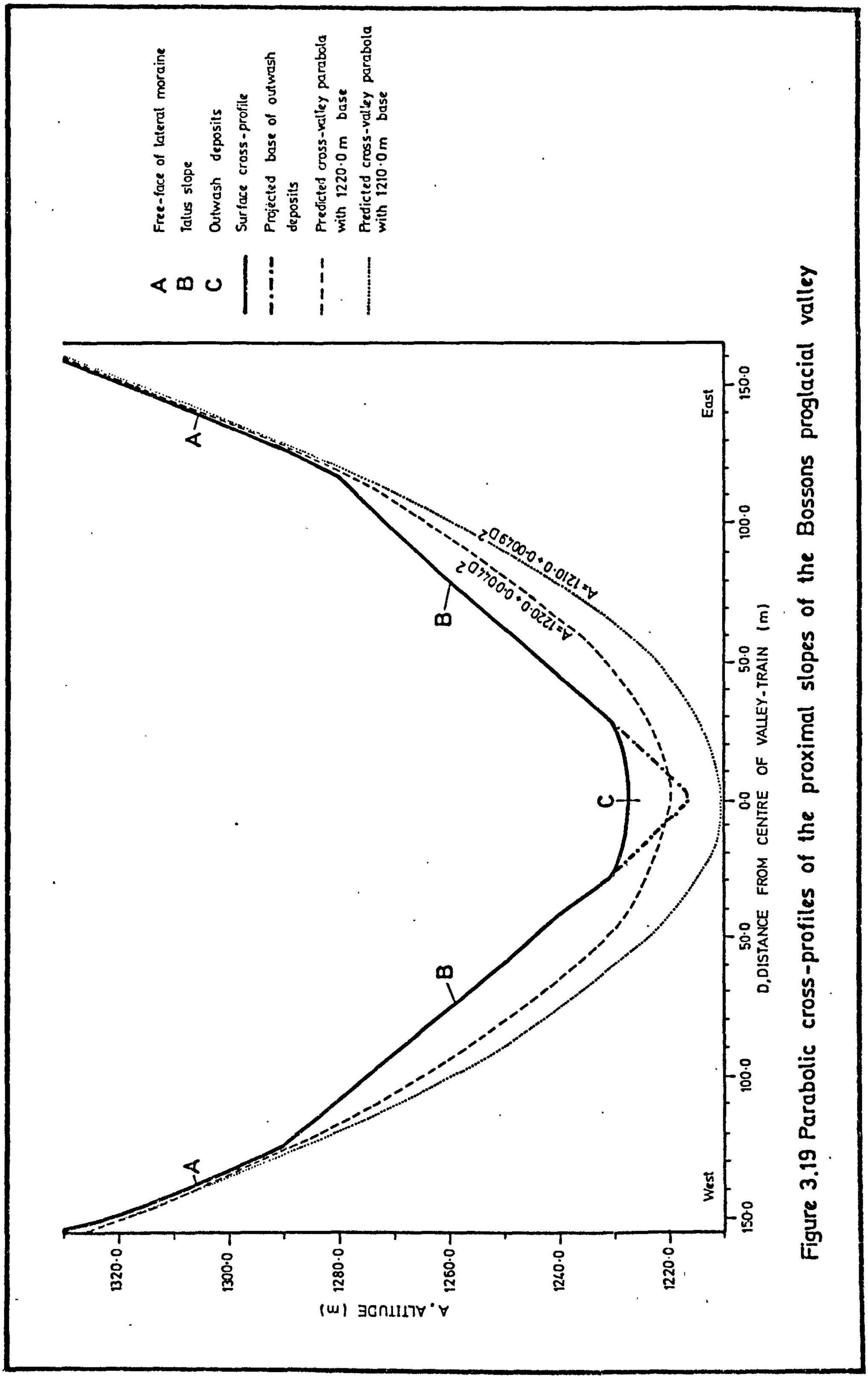
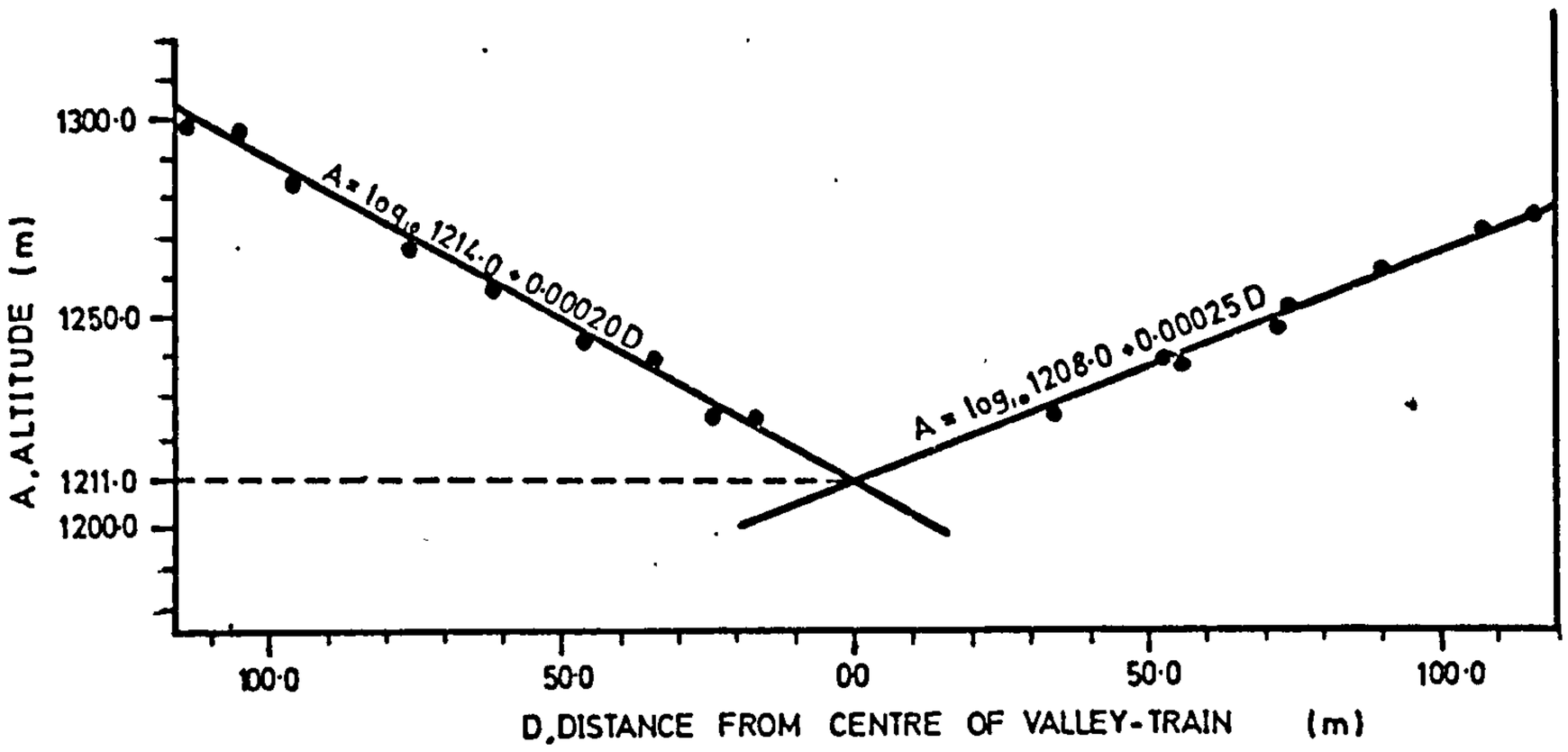


Figure 3.19 Parabolic cross-profiles of the proximal slopes of the Bossons proglacial valley

Figure 3.20 Semilogarithmic cross-profile of the proximal slopes of the Bossons proglacial valley



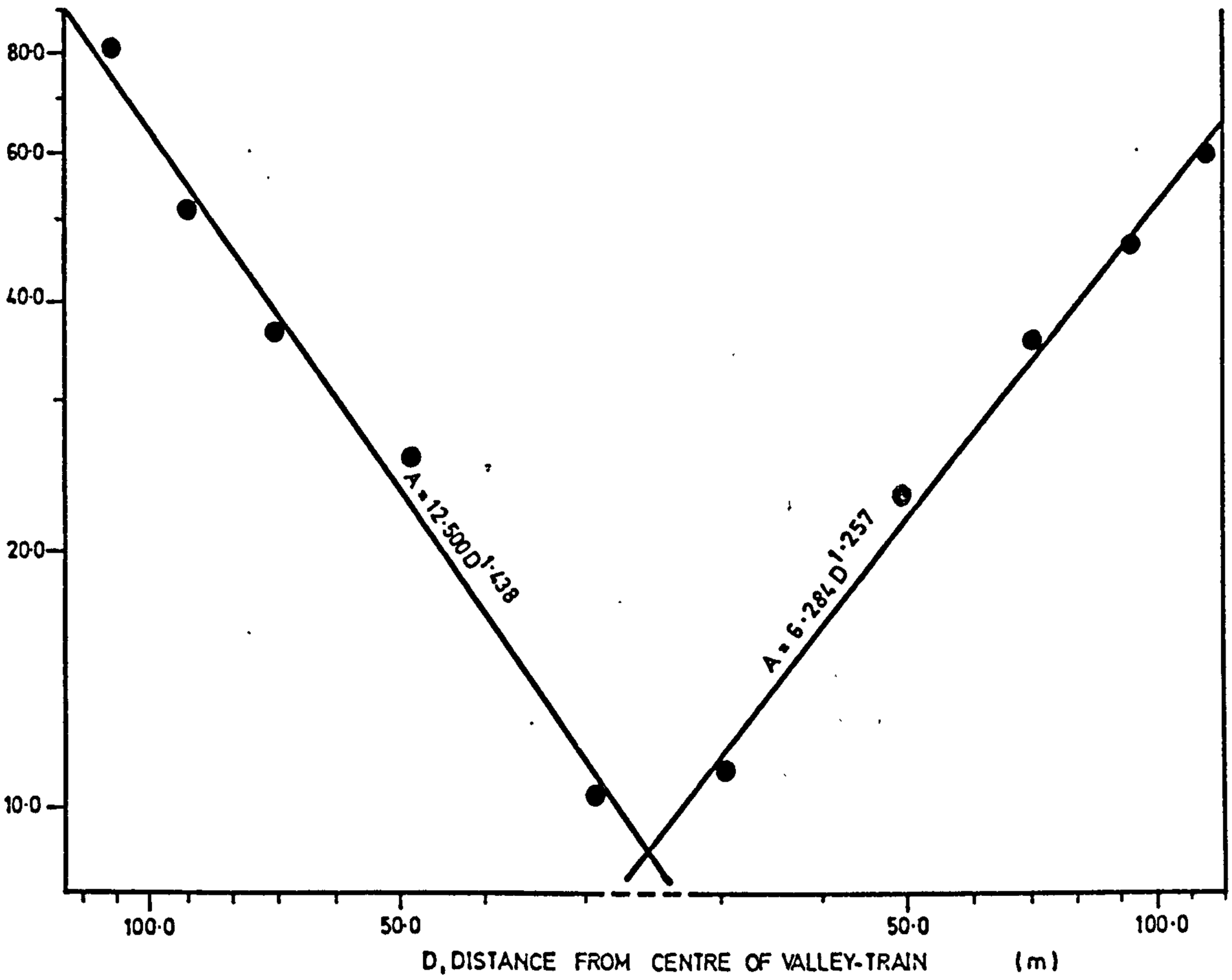


Figure 3.21 Double-logarithmic cross-profile of the proximal slopes of the Bossons proglacial valley

rock level are nearer to the value of 2 than those based on a 1210m rock level suggests that the former may be a closer approximation to an original parabolic valley cross-profile. Values of Graf's form ratio, F_R (where F_R = valley depth/width of the valley at the top of the valley sides (Graf, 1970) range from 0.275 to 0.353, and closely correspond to Graf's own measurements of glacially eroded valleys.

Calculations of profiles at downstream sites are complicated by the extensive development of talus slopes. Nevertheless, the calculation of probable thicknesses of outwash sediment and the integration of several cross-sectional areas of the valley-train deposits, mean that the total volume of outwash sediment can be estimated. The volume of proglacial valley-train deposits in July 1971 is thus estimated to have been $4.024 \times 10^4 m^3$. The mean annual rate of outwash deposition, assuming that the proglacial sediments have accumulated since ice retreated from the valley-train in 1940 (chapter 2.6) has therefore been about $1.3 \times 10^3 m^3$, representing an average annual net increase in sediment thickness of $0.531 m/m^2$. Comparison with rates of sediment removal by meltwaters may give an indication of the rates of glacial erosion (see discussion in chapter 5.5).

3.5. THE WATERTABLE

3.5.1. THE DEPTH OF THE WATERTABLE

The depth of the watertable was measured in order to determine the shape of the long profile of the watertable in relation to the long profile of the valley-train surface, and also to see if the watertable varied in height during the day. Diurnal variations would reflect the relative mobility of the watertable, indicating that seasonal variations in its height must also occur, following the annual cycle in the supply of water. The depth below the surface at which the watertable occurred was measured by digging pits at over 60 points over the valley-train in April 1974. The watertable was taken as the surface of the pool of water which formed at the bottom of the pit, and measured at least half an hour after the pit had been excavated.

The watertable was found to follow the overall gradient of the longitudinal profile of the valley-train (figure 3.22, page 150), but certain local irregularities were also apparent. Firstly, the watertable forms a depression beneath the proximal debris fan, and falls sharply from the ice-margin. Groundwater in this area is clearly derived directly from seepage of meltwaters from beneath the ice-snout. A rise in the watertable occurs where several sources of meltwater converge at the foot of the debris fan, in the 'transition zone'. The transition zone, lying 25 to 30m from the ice-margin, represents a form of 'spring line', and water can be observed welling out at the surface under hydrostatic pressure. Surface meltwaters are also more extensive, and contribute to the development of a well-saturated surface zone.

The second series of irregularities in the level of the watertable are associated with the 'abandoned' terminal moraine, which is situated at about 60m from the ice-margin. The watertable is consistently higher upstream than downstream of the moraine. Groundwater appears to become "ponded" upstream as groundwater flow is slowed down by the subsurface obstruction. Groundwater gradually seeps round the western and eastern margins of the partially buried moraine and proceeds to filter inwards towards the centre of the valley-train. Seepage is relatively slow, so that a "dry tail" is formed downstream of the moraine. The influence of the buried moraine on groundwater flow and thence on the level of the watertable extends for over 30m downstream of the abandoned terminal moraine. Seven zones of subsurface drainage have been identified and are indicated on figure 3.22 (page 150).

The level of the watertable along a profile west to east across the valley-train, taken at 50m from the ice-margin, also follows the overall rise in elevation towards the centre of the valley-train (figure 3.22). The numerous small changes in micro-relief, however, tend to mask the otherwise relatively smooth profile of the watertable, although locally, the watertable can be higher than a nearby stream.

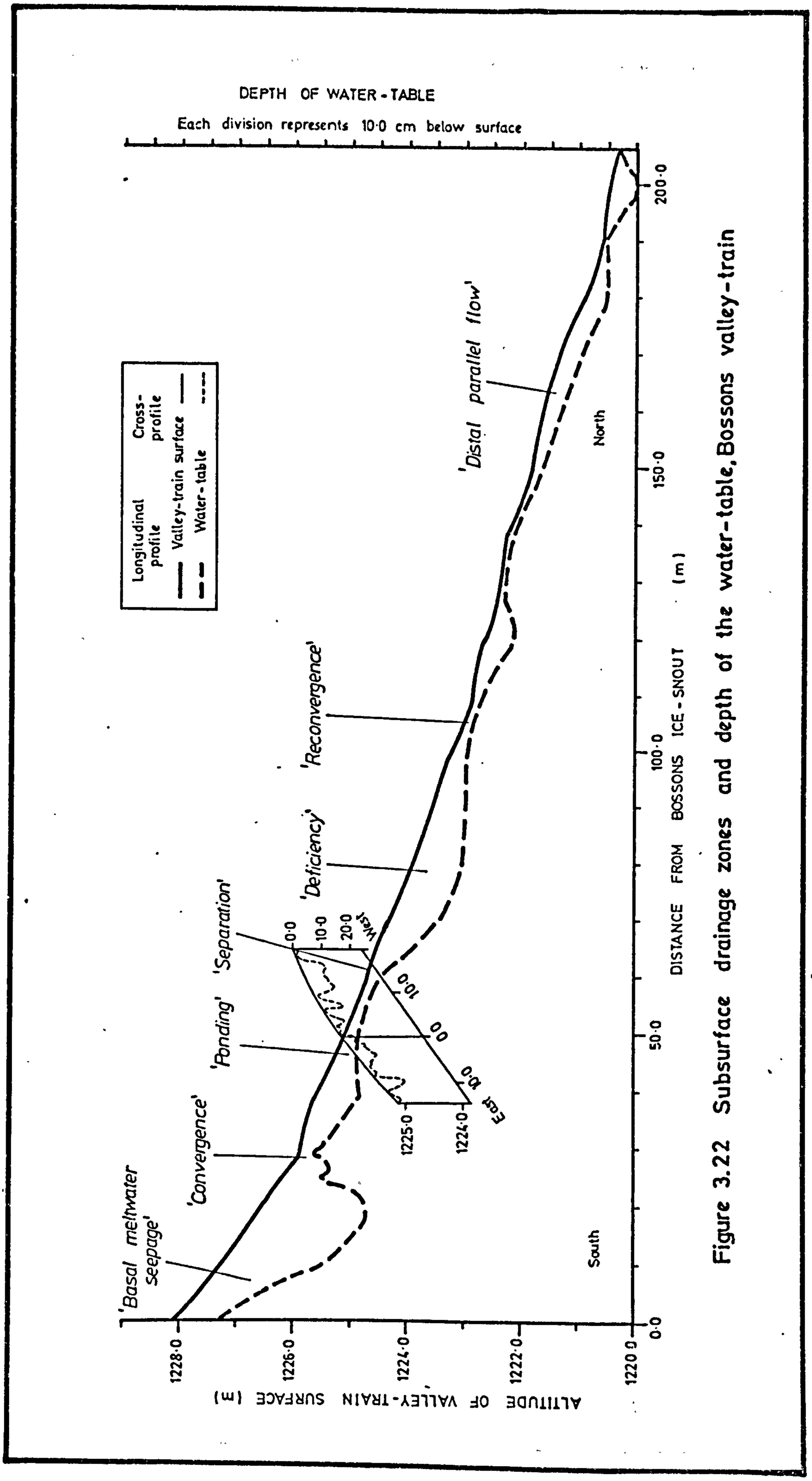
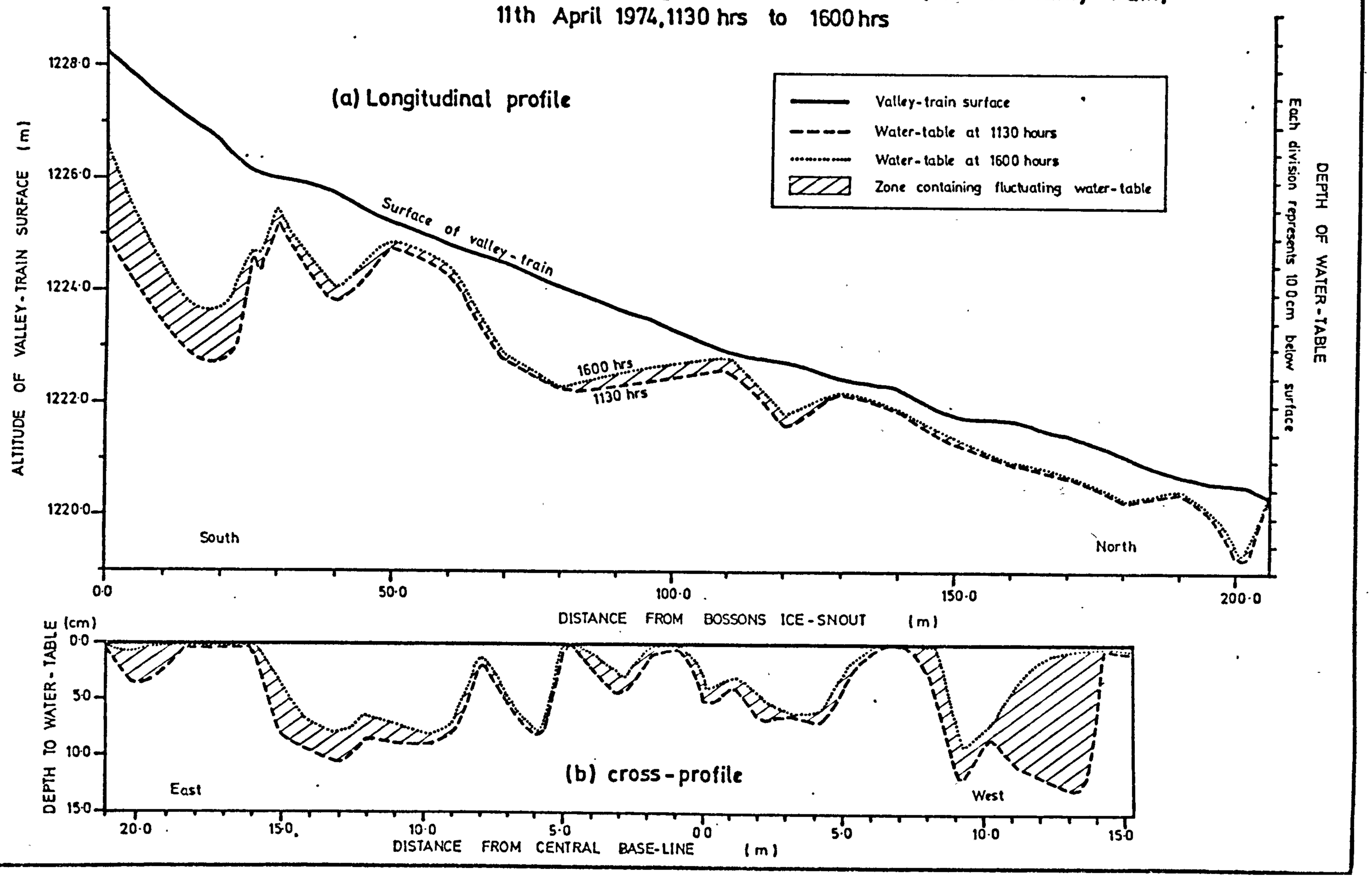


Figure 3.22 Subsurface drainage zones and depth of the water-table, Bossons valley-train

Figure 3.23 Diurnal variations in the height of the water-table, Bossons valley-train, 11th April 1974, 1130 hrs to 1600 hrs



The level of the watertable increased steadily during the day. A maximum rise of 12.5cm was recorded over the 4½-hour period. Greatest increases in the level of the watertable were recorded close to the ice-margin (see also Churski, 1973), as basal ice melting proceeded rapidly during the afternoon, and on the east side of the 'ponding' zone, prior to subsurface drainage 'separation'. The level of the watertable downstream of the abandoned terminal moraine, however, remained constant (figures 3.23a and b, page 151). Lack of pronounced diurnal variations in the watertable in the zone of 'subsurface drainage deficiency' supports the proposal that groundwater infiltration from the valley margins is very slow. The watertable in this 'dry tail' zone is both low and static, despite nearby surface meltwater streams. Thus in some areas, stream level and watertable level appear to be linked only indirectly. This suggests that the level of the watertable can depend both on infiltration of groundwater from upstream and adjacent areas, and from seepage from surface meltwaters.

The watertable is a relatively mobile feature and generally varies directly with water levels in the meltwater streams. The most dramatic evidence of a rapid rise in the water level was observed in a kettle-hole where the water level rose by over 80cm in a 5-hour period at the start of the meltseason in April 1973. The watertable, too, must exhibit similarly large variations throughout the year.

3.5.2. GROUNDWATER FLOW

Groundwaters are widely known to contribute to the winter run-off from glaciers, but the flow of groundwater in proglacial outwash sediments has been investigated only by Hjulström (1955)(see also Miller and Helmers, 1973). Church (1972) adapted Hjulström's methods to calculate the probable quantities of groundwater flow. Both those authors found that groundwater flow represented only about 0.4 per cent of the total meltwater discharge. The computations were based on Darcy's Law, which states that $Q = (k/u)As$ (see Campbell and Rasmusson, 1973)

where Q = the discharge to be calculated;

k = the permeability of the sediment, here taken as 9.9 darcys (Pardé, 1933; Church, 1972)

u = dynamic viscosity of the water, here taken as 672.21 (Church, 1972);

A = the area of the valley cross-section in which groundwater flow takes place (see below); and

s = the slope of the watertable (see below).

In order to estimate A, the depth to which groundwater flow extends has to be assumed. Hjulström's estimate of depth was based on the increased fining of sediments with depth until they were too fine to permit any significant groundwater flow. Church's estimate was based on the maximum depth (one metre) to which summer thaw was considered able to extend. In the Bossons valley, the maximum possible thickness of sediment containing groundwater flow varies from place to place, and throughout the year, while the watertable varies diurnally. The maximum observed depth of 'groundwater' was a pool of water at the bottom of a kettle-hole, which extended to a depth of almost 1.5m. Maximum depth of groundwater flow is therefore taken as 1.5m, and the minimum depth (winter flow) as 0.10m. The width of flow is taken as an average of 40m.

The slope of the watertable was found to be equivalent to the surface gradient of the valley-train, and the average of 0.03 was taken as the value of 's'.

Calculations show that groundwater flow may range from $0.0265m^3/sec.$ (depth = 1.5m), to $0.0018m^3/sec.$ (depth = 0.1m), representing 4.34 per cent and 0.30 per cent of the maximum recorded flow respectively. The latter figure is only slightly greater than the proportions calculated by Hjulström and Church, suggesting that the true depth of groundwater flow is less than 0.5m. However, the steeper average gradient recorded at Bossons (0.03) compared with the Hoffellssandur (0.02 to 0.009) and Ekalugad Sandur (0.006 to 0.0097) may account for the higher values at the height of the meltseason.

3.6. KETTLE-HOLES

Kettle-holes, or shallow proglacial depressions present in the outwash deposits, have been observed only at the beginning of the meltseason (Figures 3.24 and 3.25, page 155). No depressions have been observed between July and September since they become filled in with sediment (see Bogacki, 1973). The depressions always occur within the 'transition zone' at the break of slope between the proximal and distal slopes. The depressions present in April 1974 occupied almost identical positions to those present in the previous spring (see Plates 2 and 3), despite having been completely infilled with sediment during the intervening summer. The origin and subsequent re-excavation of the kettle-holes probably result from the operation of a combination of proglacial processes.

3.6.1. THE ORIGIN OF THE KETTLE-HOLES

Several theories have been proposed to account for the origin of kettle-holes. The most widely accepted explanations are associated with the melting out of buried ice, but other hypotheses have also been postulated (e.g. Clark, 1969).

(1) Melting of buried ice. Many authors (e.g. Geikie, 1894; Gilbert, 1904; Embleton and King, 1968; Fahnestock, 1969; Price, 1969, 1971; Galon, 1973b; Klimek, 1973; Francis, 1974) agree that kettle-holes result from the melting out of buried ice. The buried ice may be in the form of a stagnating ice terminus (e.g. Tarr, 1909; Gilbert, 1910), or in the form of blocks or thin sheets of ice which have become detached from the major ice body (e.g. Thwaites, 1926; Simpson, 1933; Cotton, 1942; Price, 1965; Jewtuchowicz, 1971, 1973).

Experiments were carried out by the author to determine the most likely ways in which kettle-holes might develop. A glass tank 30cm long, 20cm wide and 20cm deep, was filled with alternating layers of fine (smaller than 2mm) and coarse (larger than 2mm) outwash sediments, with fines on the surface. Blocks of ice were placed at depth or on the surface under different conditions of surface slope and saturation. Rates of ice melt, extent of moisture absorption by the sediments, and any changes in surface

Figure 3.24

A kettle-hole in winter, Bossons valley-train

Figure 3.25

A kettle-hole in the meltseason, Bossons valley-train

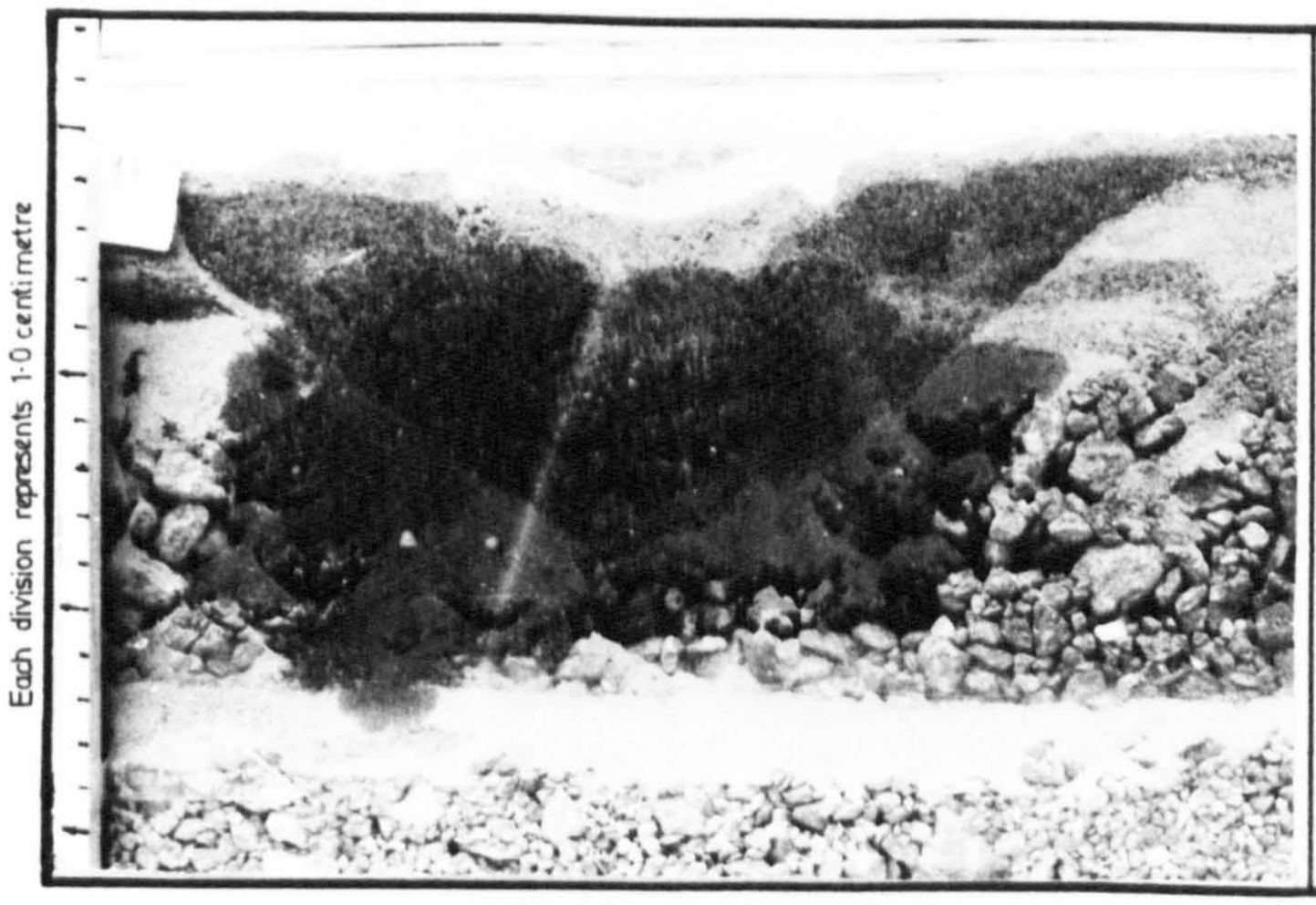
1 Kettle-hole
2 Progressive sediment infill from feed-in meltwater stream
3 Infilled kettle-hole
4 Cracking and collapse of overhanging ice-platform of Bossons glacier snout
5 Terminal moraine

relief were recorded. Further details are discussed in Maizels (in preparation, b).

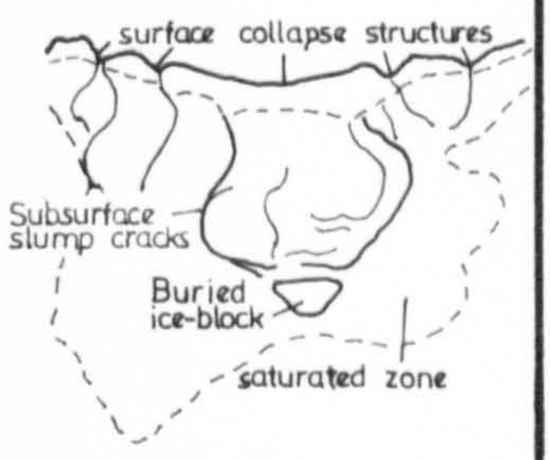
The melting of a block of buried ice produced a distinctive collapse structure, whose depth corresponded with the thickness of ice (figure 3.26, page 157), and whose shape roughly corresponded with the shape of the buried ice mass.

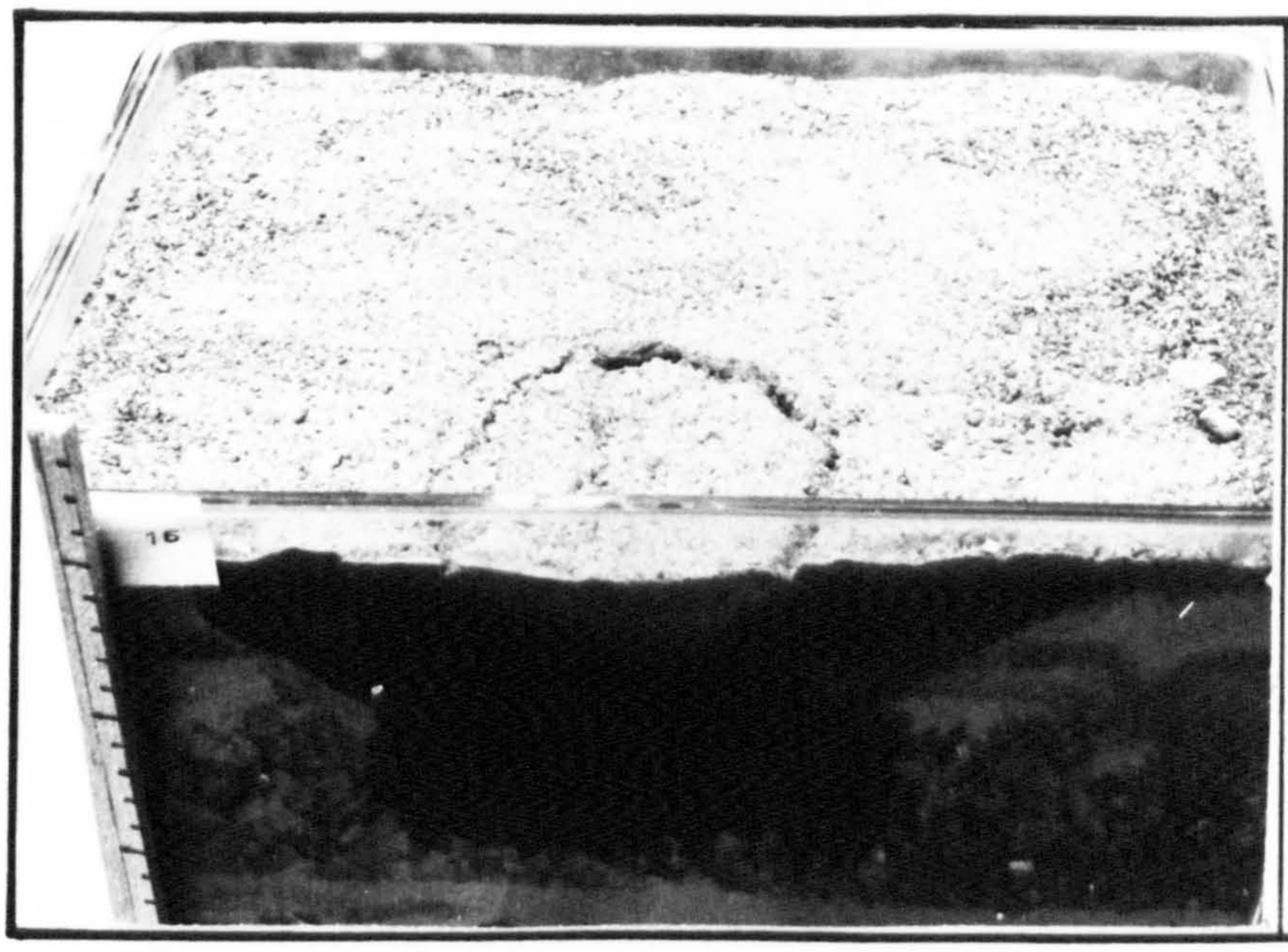
The mechanisms which separate an ice block and subsequently cause it to be buried by outwash sediments are less clear. Blocks of ice may fall off the ice-snout, as was observed at Bossons in 1967, and they may be carried a short distance downvalley during flood discharges (e.g. Hobbs, 1931; Broscoe, 1972). Thin sheets and fragments of ice may also become separated from the ice-snout if it forms overhanging 'platforms' which subsequently crack and slip on to and down the proximal slope (see section 3.1.1.). The burial of ice can be very rapid in ice-marginal locations, especially when the rate of outwash deposition is also rapid. It is more unlikely for ice blocks to be completely buried in distal areas, where rates of deposition are reduced (see below). The mechanisms which cause the ice to melt out, once it has been buried, have hardly been discussed in the literature. Gilbert (1910) suggested that groundwaters may effect melting of buried ice, and since groundwater flow has been observed filtering into the base of kettle-holes, both at Bossons and by Price (1969) at Breidamerkurjökull, the melting of buried ice by groundwater seems a feasible process. However, in many proglacial areas permafrost may be present to within a metre of the surface even during the meltseason (Church, 1942), while at Bossons interstitial ice was present in spring at depths of over 30cm below the ground surface. The groundwater which melts buried blocks of ice may therefore be derived from the melting of ground ice. Ground ice melts downwards from the surface in the melt-season, but also melts upwards from the base of the permafrost layer as a result of geothermal heat transfer. It seems likely that the melting of buried ice blocks results from both groundwater flow and from geothermal heating. Warm

FIGURE 3.26
COLLAPSE STRUCTURES PRODUCED BY THE MELTING OF BURIED ICE

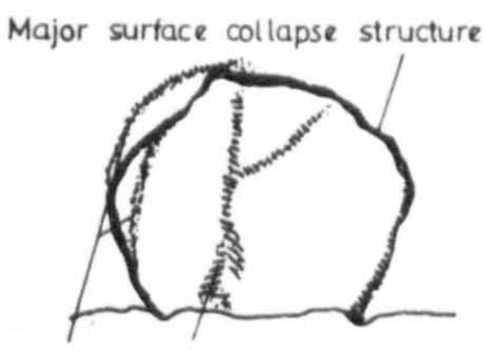


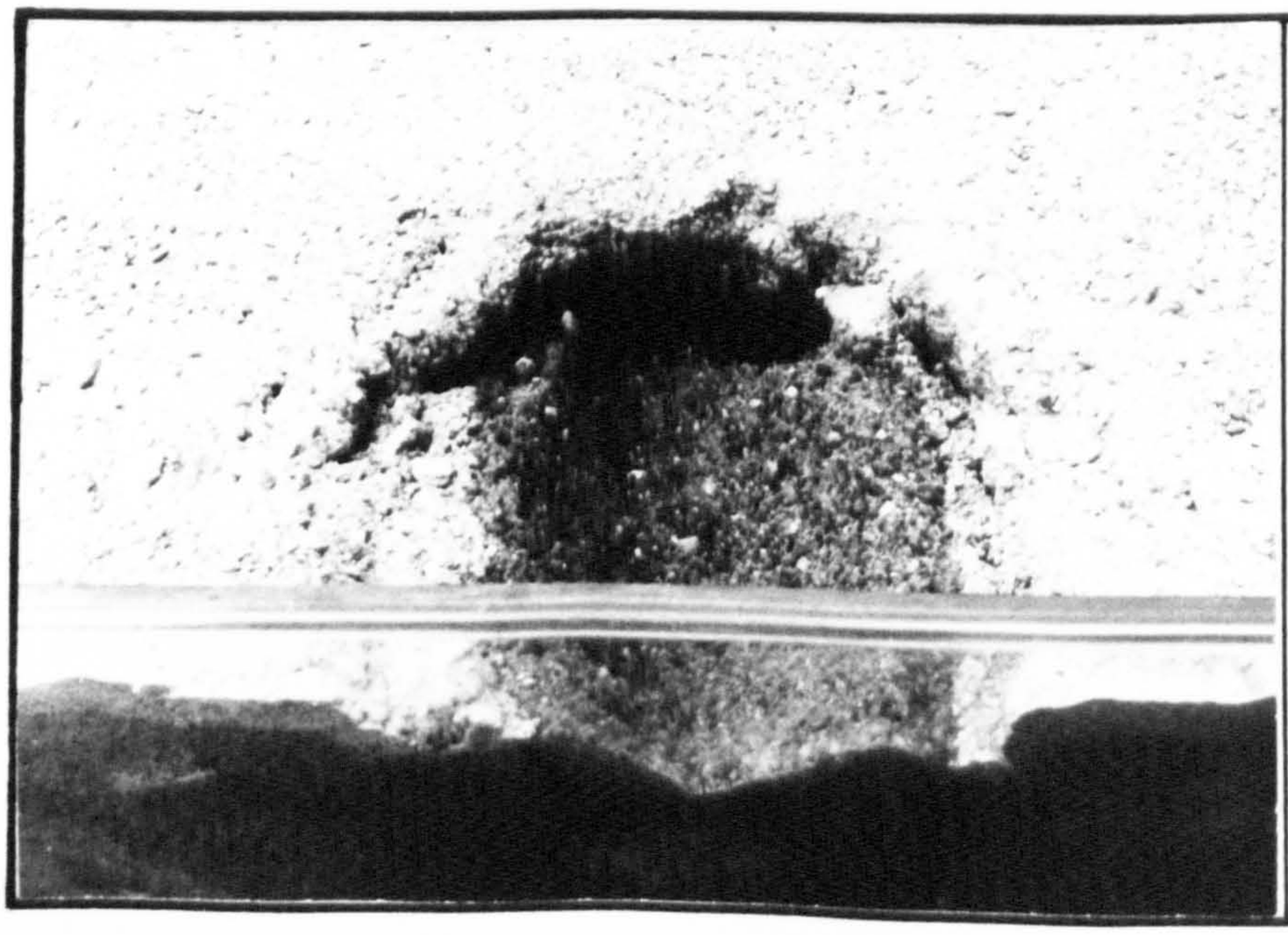
(a) cross-section



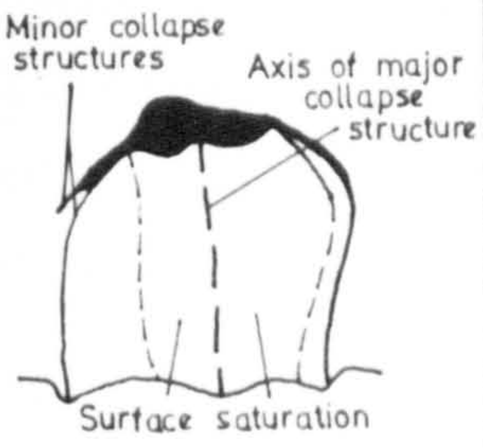


(b) in plan





(c) surface depression after final ice-melt



air and warm rain during the meltseason may also filter downwards to assist melting. The relative contribution and rate of activity of each process is unknown.

Several workers have recorded the rapid formation of kettle-holes. Francis (1974) described the development of a flat outwash plain to a complex ridge-and-pit area after a period of eight years; Price (1971) recorded a large increase in the number of kettle-holes present on Breidamerkurjökull between 1966 and 1968; Fahnestock (1963) recorded the formation of kettle-holes within a period of only 24 hours on the White River valley-train. The rapidity of formation of kettle-holes depends on the rapidity of ice-melt, and therefore on the rate of activity of geothermal heating and groundwater melting, on the thickness of buried ice, and on its proximity to the surface from which subaerial influences can penetrate.

The depth at which the ice has been buried depends on the rate of outwash deposition. The deepest kettle-hole at Bossons extended 2m below the outwash surface, suggesting that, if buried, the ice would have been at least 2m thick. The glacier was last in the locality of this kettle-hole in the mid-1940's so that the ice may well have been buried about 20 years ago. Stagnant, detached masses of ice may have accumulated upstream of the abandoned terminal moraine, in the region of 'ponded groundwater drainage', in which rapid sediment aggradation occurred. However, measurements of the changes in elevation of the longitudinal profile indicate that over 2m of aggradation took place in this area between 1971 and 1974, suggesting that the kettle-holes were initially much deeper.

(2) <u>Melting of partially buried ice</u>. Blocks of ice transported by meltwaters on to the outwash plain may become partially surrounded by fluvioglacial sediments. Insolation from above, undercutting by meltwaters below, and gravity sinking into saturated sediments result in the ice-block melting before it is completely buried, subsequently producing a depression (e.g. Davis, 1890; Thwaites, 1926; Wright, 1937; Lougee, 1940; Cotton,

1942; Michel, 1969; Klimek, 1973; Paterson, 1974).

The author's experiments showed, however, that a high rate of sediment deposition by meltwaters flowing within the channel in which the ice block had been dumped, almost obliterated any incipient kettle-hole so that only a faint depression could be detected once the ice had finally melted (figure 3.27, page 160). A circular depression was also produced in saturated sediments drained by 'sheet flows' of water (figure 3.28, page 161). The depression was also rapidly infilled with sediment. This process of kettle-hole formation only seems likely to occur in areas of relatively low meltwater flow, with little sediment transport after the initial dumping of the ice-block. The kettle-holes at Bossons are not likely to have resulted from partial burial of ice blocks, firstly because meltwater discharge is not sufficiently powerful to transport large blocks of ice, and secondly because coarse proglacial sedimentation is so rapid.

(3) Melting of ice lying on the surface

Blocks or sheets of ice which have become detached from the active, retreating ice-mass, may melt out in situ on the outwash surface before being affected by proglacial sedimentation. Several workers claim that kettle-holes result from in situ melting of surface ice (e.g. Mannerfelt, 1945; Ives and Andrews, 1963; Broscoe, 1972). The blocks of ice may sink into the outwash surface if they are sufficiently heavy and if the substratum is sufficiently unconsolidated. This mode of formation is accounted for by the sapping of the base of the ice block and of surrounding fine sediments by meltwater from the ice block itself. The fine sediments then filter through the outwash deposits carried by groundwater. Eventually, resettling and sediment sapping act to create a pit-like feature.

The experiments indicated that, unless the ice block is exceptionally heavy, no depression results from the surface melting of an ice-block (figure 3.29, pages 162 and 163). Meltwater filters downwards into the underlying sediments as the block of ice melts, and sediment particles

FIGURE 3.27
MELTING OF DEBRIS-COVERED ICE-BLOCK IN MELT-WATER STREAM CHANNEL

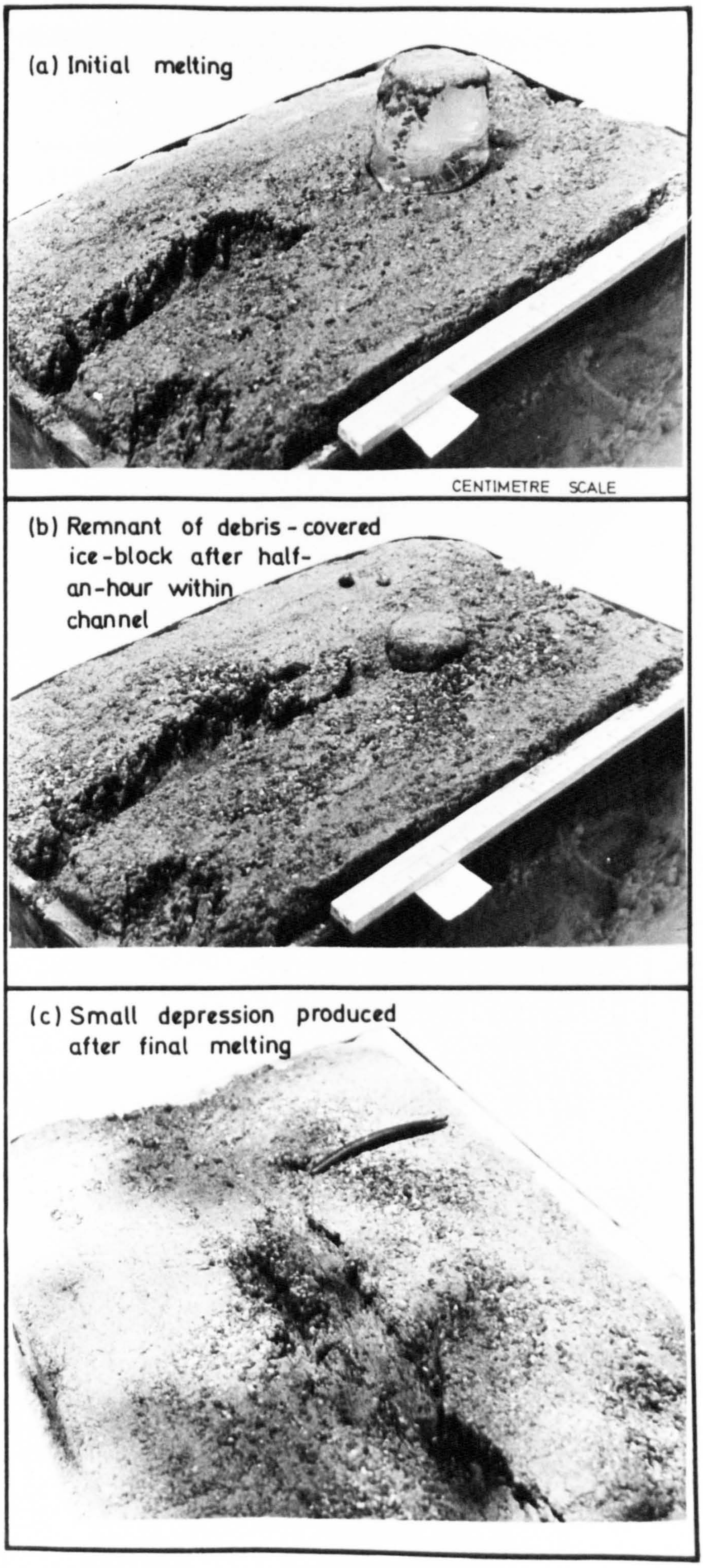


FIGURE 3.28
MELTING OF AN ICE-BLOCK ON A SATURATED, SLOPING GRAVEL SURFACE

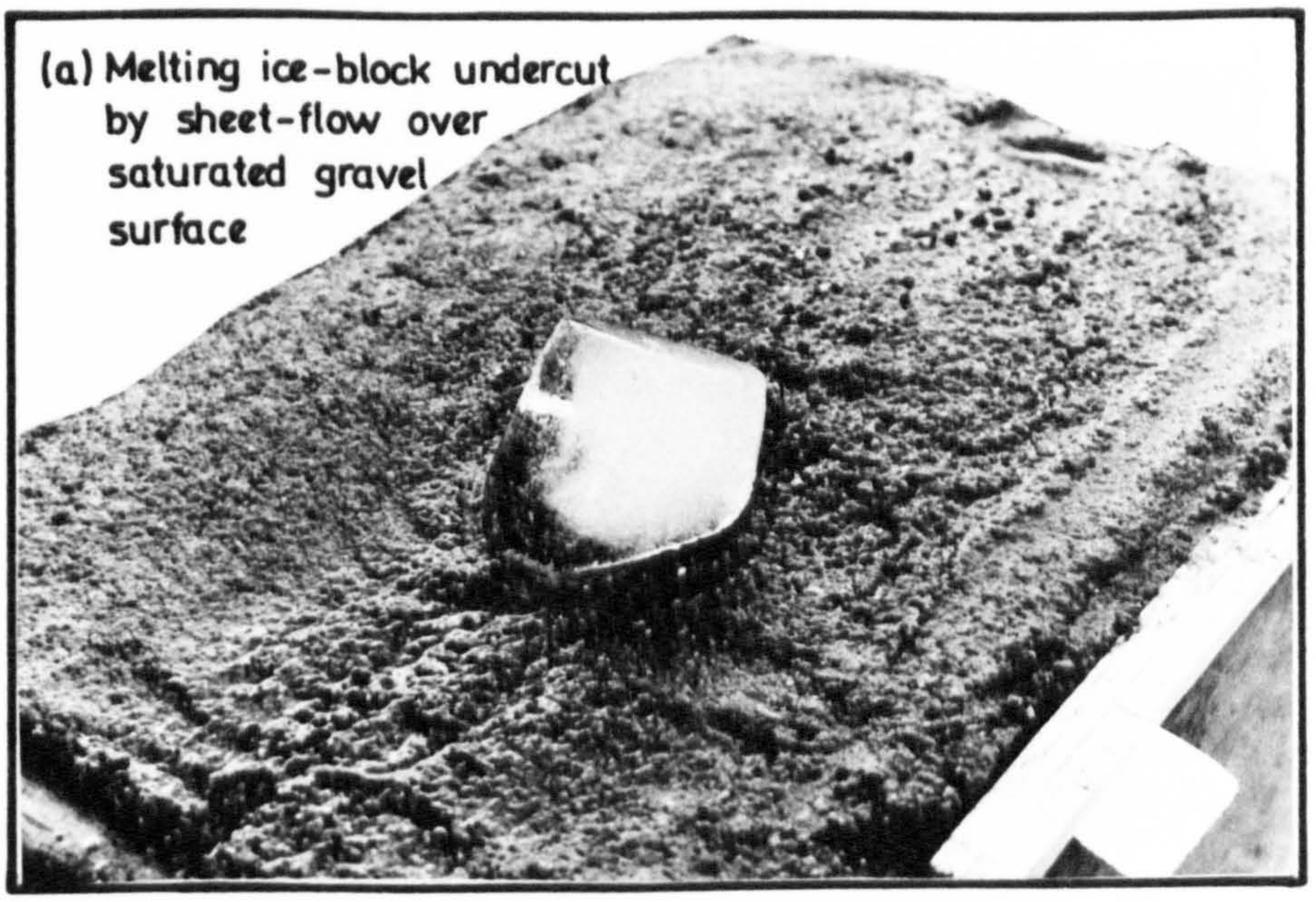
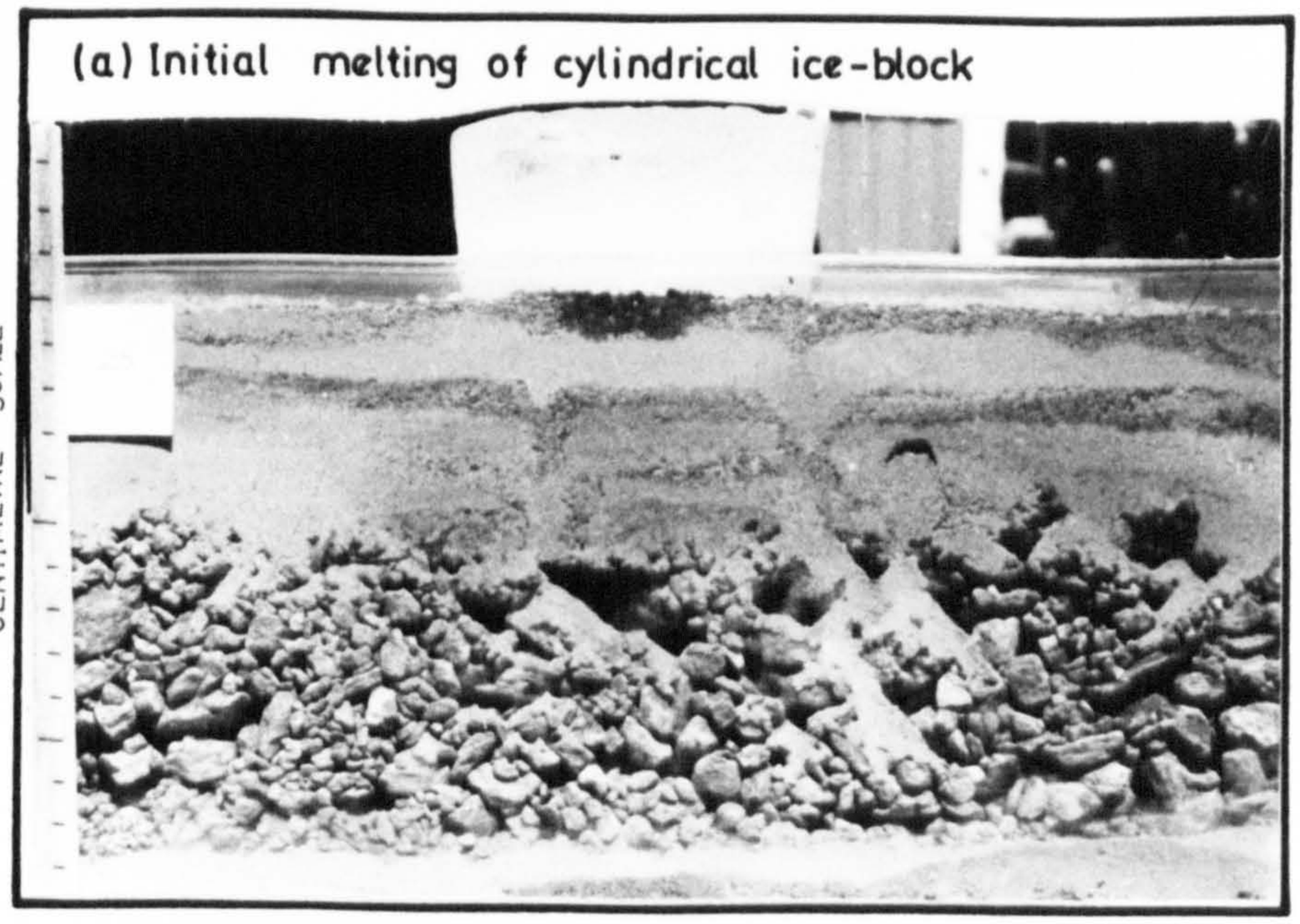


FIGURE 3.29
MELTING OF AN ICE-BLOCK ON A DRY GRAVEL SURFACE



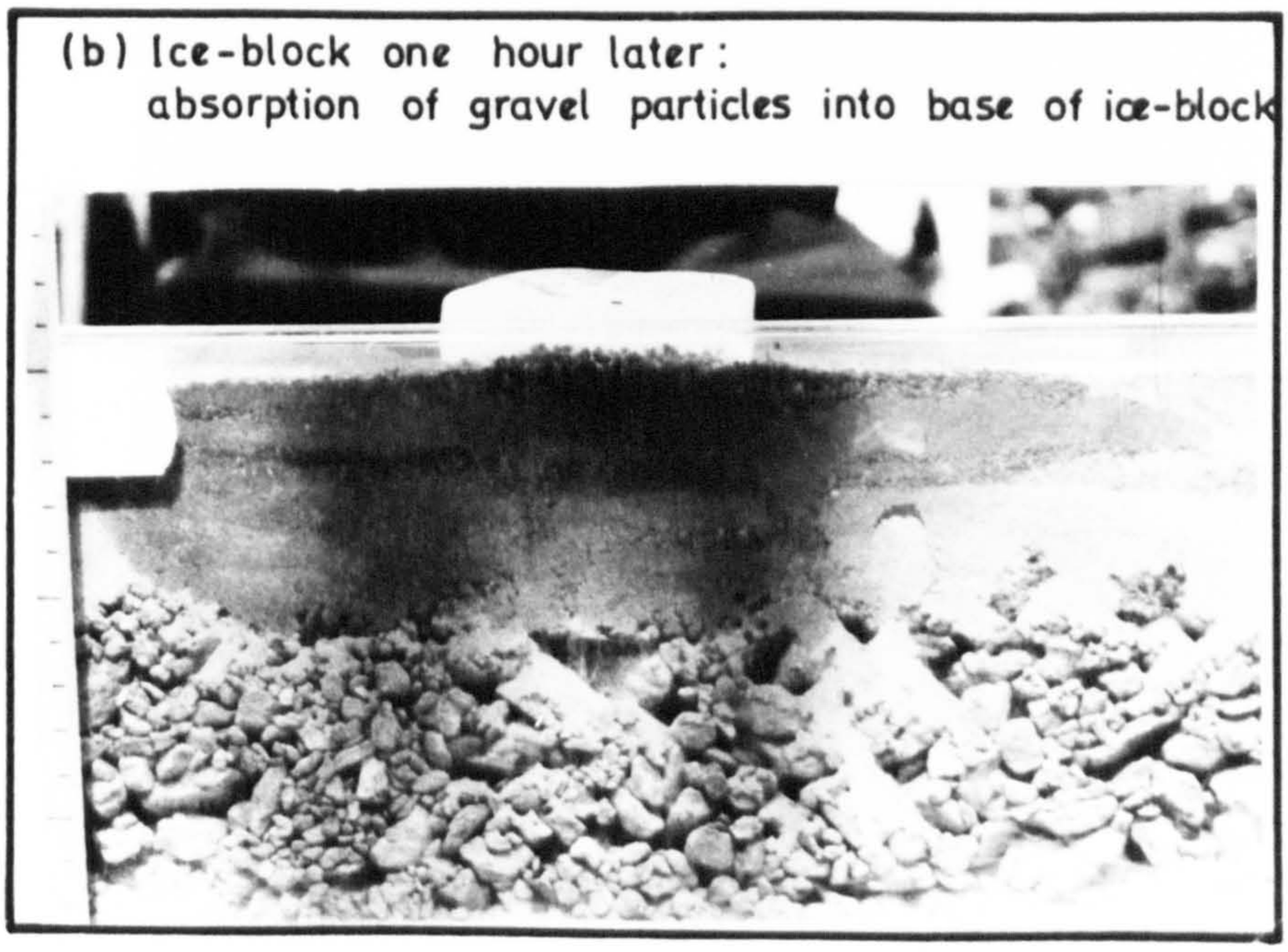


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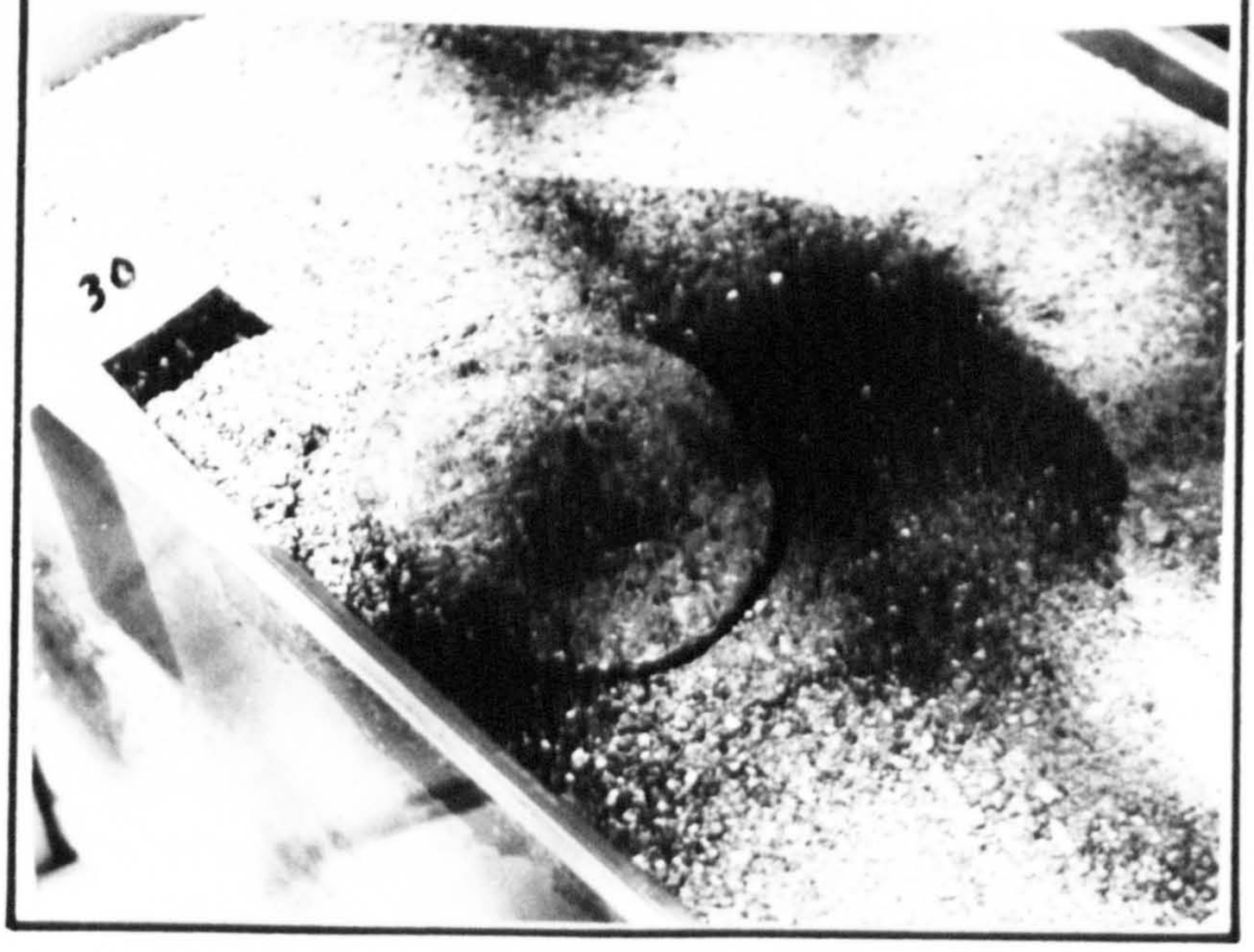
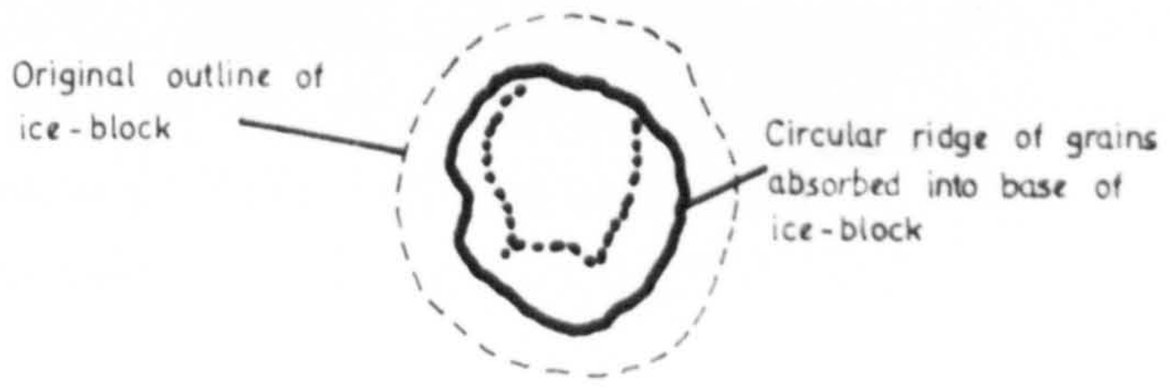
FIGURE 3.29 (continued)

(c) Final stage of ice-melt



(d) Circular ridge of grains left on surface after final melting of ice-block



from the outwash surface are drawn up into the basal layer of ice (figure 3.29b). A series of ridges of sediment are left on the outwash surface when the ice has melted, the ridges forming a pattern which corresponds to the outline of the original block of ice (figure 3.29d). Thus, rather than producing a depression, this form of ice-melt creates a pattern of surface relief.

(4) Other modes of formation

A number of other theories have been proposed to account for the origin of proglacial pits. Thwaites (1926) suggests that stream scour or aeolian deflation might produce surface depressions. Localized stream scouring is probably partly dependent on a change in the nature of the channel, the channel sediment, and channel slope. The well-saturated 'transition zone' at Bossons may, for example, allow rapid scouring to occur during periods of high or increasing meltwater discharge to produce proglacial depressions. Aeolian deflation, however, is probably precluded by the coarseness and dampness of the surface of the outwash deposits. Embleton and King (1968) state that proglacial basins might originate from the freezing of small lakes, meltwater streams or ground ice. The 'small lakes' may themselves be kettle-holes, while the influence of the freezing of surface or ground ice on kettle-hole formation is not clear. Flint (1971) describes depressions which develop by differential thawing of ground ice and through irregular deposition of moraine at the ice-margin or across pre-existing valleys. The latter origin of proglacial pits cannot account for the Bossons kettle-holes.

3.6.2. The annual 'cycle' of kettle-hole development

The development of the proglacial depressions appears to follow a cyclical pattern, once the kettle-hole has initially formed. The depression is either completely dry, or floored by a shallow, frozen pool before the

meltseason commences (figure 3.24, page 155). The pit rapidly fills with water as soon as melting begins; one of the Bossons kettle-holes took only six hours to fill and then to overflow into outlet channels (figure 3.25, page 155). Fine sediments are gradually washed into the pool, producing minor delta formations and varve-like bottom sediments or rhythmites (e.g. Brunnacker, 1975), comprising silts and clays, as the meltseason progresses. The infilling stage is completed when the kettle-hole is completely filled with fine sediment, intercalated with coarse debris from collapsing walls of the kettle-hole, and overlain by coarse material contiguous with adjacent channels, so that the site of the depression can no longer be distinguished (figure 3.5, page 119) (see also Churski, 1973).

The infill is subsequently restructured or re-excavated towards the end of the meltseason so that by the following spring the proglacial depression is again present at the same site. Re-collapse of the infill may occur if the bottom sediments were initially laid down on the surface of the frozen pool, before it had time to thaw. However, if the bottom ice was in contact with meltwaters filling the pool, most of the ice at least is likely to have melted. Several of the kettle-holes were not floored by ice but by mud, so that repeated annual collapse of sediment infill is not a likely cause of kettle-hole persistence from year to year.

Groundwater seeping through the sediment infill into the surrounding outwash deposits or into the underlying deposits, as a form of subsurface piping, may remove much fine sediment, and this may contribute to initial resettling of sediment and lowering of the sediment surface towards the end of the meltseason, at the same time as the watertable is also being lowered. Removal of most of the sediment infill, and progressive lowering of the ground watertable, both act to create and maintain a dry proglacial depression, unless it is sufficiently deep to be floored by a shallow, frozen pool of water, until the start of the

following meltseason. Galon (1973a) describes a kettle-hole on Skeidararsandur which may have resulted from 'former artesian flow gushing out from under the glacier margin'. Upwelling of groundwater at the break of slope on the Bossons valley-train may repeatedly produce a proglacial depression of this kind.

3.6.3. Kettle-hole morphology

Kettle-holes are often asymmetric, and with the steepest sides varying in aspect in different sites. Asymmetry appears to result from local stream erosion, undercutting, scour and 'headward' erosion upvalley. Filtration of groundwater and meltwater from lying snow also causes resettling and slumping of kettle-hole walls as the fine sediments are washed out. The kettle-holes are almost circular, but asymmetry is thus accentuated by steeper, shorter slopes and extension of kettle-hole walls downstream to form channel banks. The depth of kettle-holes is difficult to calculate, because the depth of water and thickness of sediment infill are continually varying. Thirty-six cm of fine sediments were recorded at the bottom of one kettle-hole in April 1973, while the depth of the surface of the sediment below the valley-train surface was 1.4m. The kettle-hole would have been infilled with a total of 1.76m of sediment by June or July at the latest. Sediment infill is derived from three main sources: (a) Inflowing surface meltwaters from ice and snow contribute fine sediments to form a 'varve-like' lacustrine deposit, partly diurnal in nature. When infilling is almost complete, coarse fluvioglacial deposits begin to appear; (b) Groundwater also contributes very fine sediment from beneath and within the bottom sediments and from the kettle-hole walls; (c) The kettle-hole walls themselves become undercut and cave in to form saturated slump deposits on the floor of the kettle-hole.

Chapter 4

THE SCOTTISH OUTWASH DEPOSITS:

THE PHYSICAL BACKGROUND

Chapter 4

THE SCOTTISH OUTWASH DEPOSITS: THE PHYSICAL BACKGROUND

4.1. INTRODUCTION

The locations of the North Esk, the Westwater, the South Esk and the Glen Lethnot outwash deposits have been outlined in Chapter 1 (and see figure 1.2, page 44). It was also stated in Chapter 1 (section 1.4.2.) that these outwash deposits were selected for study because their approximate dates of formation are known (i.e. Late Weichselian), they comprise schist and granite lithologies, which also occur at Bossons, and they contain a relatively large number of exposures. The North and South Esk outwash deposits are the most extensive and comprise the terraced remnants of a former 'sandur' plain (figure 4.1, page 169); the Westwater outwash forms an important tributary deposit of the North Esk outwash plain. The Glen Lethnot deposits are confined within a narrow upland valley and have been dissected to a depth of several metres so that shallow terrace remnants of a former valley-train now remain (figure 4.2, page 169).

The observation of ancient proglacial environments is limited to the morphology and sedimentology of the outwash deposit, whereas study of the present-day proglacial valley of the Bossons glacier has been able to include some of the physical aspects of the environment such as climate, short-term glacial history, hydrology and biogeography. The comparison of paleo-outwash deposits with a contemporary proglacial environment can, nevertheless, provide valuable information about relations between particle lithology, size and shape, sedimentary structures, outwash morphology and slope. Differences between observations of several sets of outwash deposits inevitably reflect, in varying degrees, local and regional geomorphological conditions. Interpretation of the sediment variations can only be adequately approached through investigation of some of the features of the physical environment associated with the formation of

Figure 4.1
The North Esk gravel pit, looking northwards to Glen Esk

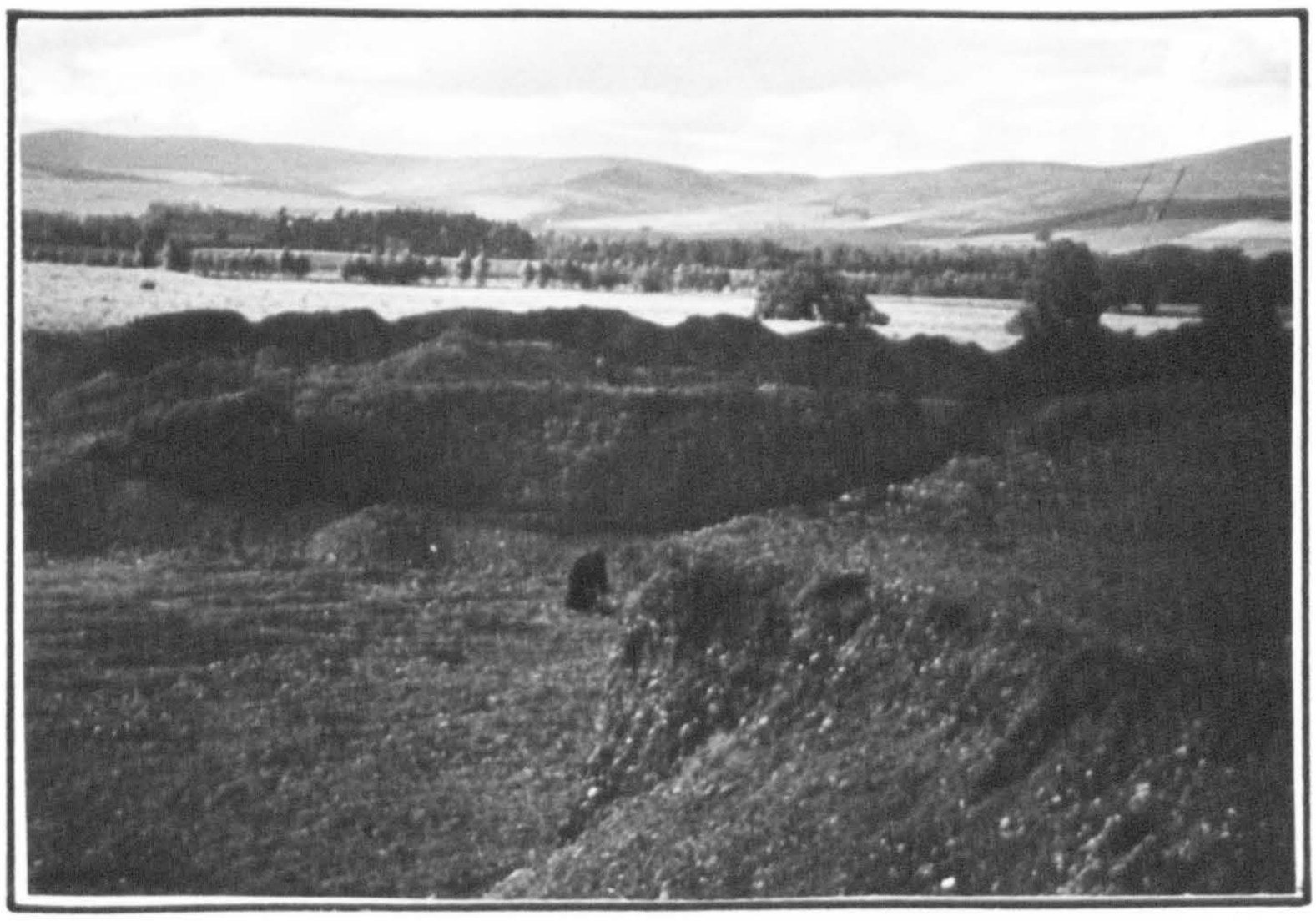
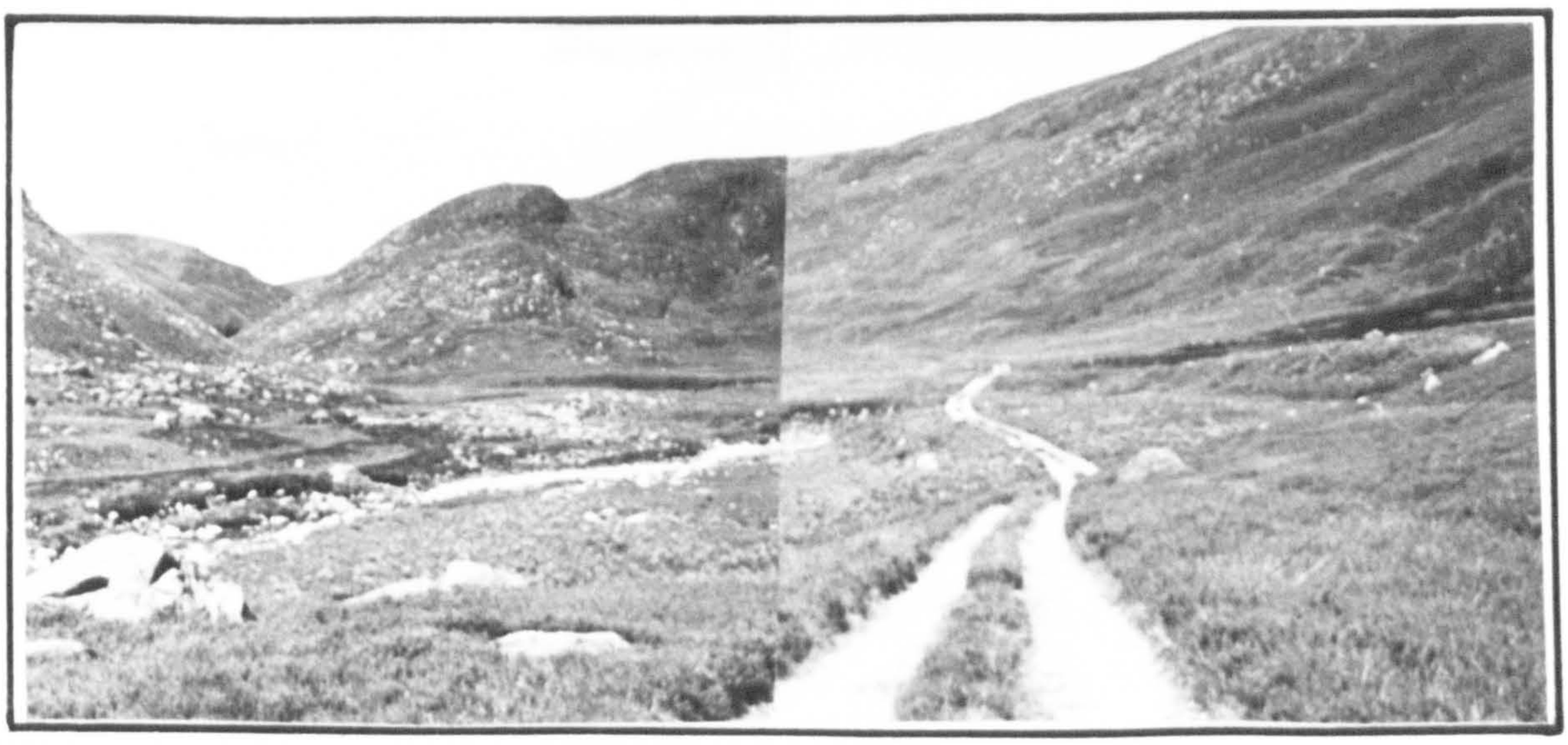


Figure 4.2
The head of Glen Lethnot, looking upstream



each deposit. These features include the lithological composition, scale and age of the deposit, the glacial chronology of the area, and the structural and sedimentological deformations effected after initial deposition. An additional problem in determining the physical background to each set of deposits is the availability of exposures for observation and sampling of the outwash sediments.

Few of these problems have been discussed per se in the literature, even though many other workers have attempted to compare certain features of fluvial deposits (e.g. Illies, 1949; Schlee, 1957b; Pelletier, 1958; Pettijohn, 1962; Martini and Ostler, 1973; Eynon and Walker, 1974; and see Chapter 1). Some workers have drawn attention to differences in the geological and climatological background to alluviation (e.g. Church and Ryder, 1972) and to sampling problems (e.g. Krumbein, 1959a).

The physical background to the four series of Scottish outwash deposits is presented in sections 4.2 to 4.7. Discussions also include examples of how the difficulties involved in comparative studies of outwash deposits may be tackled. The problems of sampling the paleo- and the contemporary outwash deposits are considered in some detail in section 4.8.

4.2. GEOLOGY OF THE MELTWATER BASINS AND LITHOLOGICAL COMPOSITION OF THE OUTWASH DEPOSITS

4.2.1. INTRODUCTION

The different series of outwash deposits selected for comparison should be composed, as far as possible, of similar lithologies, in order to simplify interpretation of sediment characteristics, and even of morphological features such as stream gradient, which may be related to particle resistance (Chapter 7). Many workers have omitted to mention the lithology (or lithologies) of the sediments they are studying or have not regarded differences in lithology within or between different deposits as important (e.g. Hjulström, 1952; Jewtuchowicz, 1955,

1962a,b, 1971; Krumbein and Lieblein, 1956; Fahnestock, 1963,1969; Williams and Rust, 1969; McDonald and Banerjee, 1971; Boothroyd, 1972). However, for valid comparisons between different deposits the proportion of each lithological group should be stated, and particle measurements should be related to separate rock types, for each lithology responds in a different way to the processes operating (e.g. Hövermann and Poser, 1951; Cailleux and Michel, 1967; Ehrlich and Davies, 1968; Bradley et al., 1972; Stanley and Wayne, 1972). The results of the analysis of each individual rock type must finally be integrated to represent the complete mixture forming the whole, heterogeneous deposit.

4.2.2. GEOLOGY OF THE SCOTTISH RIVER BASINS AND MELTWATER OUTWASH DEPOSITS

The Esk and Westwater outwash deposits were selected for study basically because they include lithologies similar to those at Bossons (see Chapter 1). Schists and granites are the dominant rock types present in the Bossons outwash deposits, while in the Strathmore deposits schist forms the major proportion of a suite of lithological groups (see figure 6.14, page 347) derived from the whole drainage basin (see figures 4.3a and b, pages 174 and 175). Nine lithological groups have been differentiated in the Strathmore deposits[1]:

1 Schists
2 Granites
3 Gneisses
4 Quartzes and quartzites
5 Altered Grits
6 Dolerites
7 Lavas
8 Sandstones
9 Others (e.g. limestone, conglomerate, serpentine).

---

1. The classification of some specimens was complicated by the apparent continuum existing between several lithological groups, such as granite - gneiss and coarse sandstone - altered grit - quartzite - quartz. Similarly, some 'dolerite' clasts were difficult to differentiate from 'lava' clasts, and for this reason, groups 6 and 7 are combined for some analyses. Replicate counts indicate that 3.35 per cent of coarse particles may be incorrectly classified.

Differences in the composition of the constituents of each lithological group (e.g. Group 1 comprises 'schist', 'mica-schist', 'hornblende-schist', 'quartz-schist', 'quartz-mica-schist', and 'gneissose-schist') may account for some of the variation in results obtained.

Data relating to the schistose and granitic lithologies are thus more directly comparable between the Bossons deposits and the Strathmore deposits than if totally dissimilar lithologies were present. Data presented later (Chapter 6) suggest that each lithology is structurally similar in each area. The wider range of lithologies present in the paleo-deposits also allows closer examination of the effects of transport on different rock types, and the possible distances of transport from a variety of source localities.

4.2.2.1. Solid Geology of the North Esk and Westwater basins

(1) The Upper reaches of the North Esk and Westwater drainage basins lie within the Highland region, and the major watersheds follow a series of Grampian peaks rising to over 700m (approx.) OD. The upper reaches of the North Esk basin, covering an area of about 420 km^2, are composed of Dalradian schists, gneisses, granites, quartz, quartzites and altered grits, penetrated by dyke-like intrusions of dolerite and quartz-porphyry, and by isolated outcrops of limestone (figure 3.1a).

(2) The Highland Boundary Fault Zone consists of a series of parallel, southwest-northeast trending lithological zones. The Fault Zone extends for about 1.5 km along the present valley of the River North Esk. The geological zones which outcrop from north-west to south-east (i.e. in a downstream direction) pass from Dalradian schists and wedges of the shales and limestone of the Margie series to the Jasper and Greenrock series, and then to the grey, gritty, quartzitic sandstones of the Margie series. Two dolerite dykes cut the Margie series, and are aligned parallel with the Highland Boundary Fault (Johnstone, 1966). The acid intrusion of Lintrathen Porphyry occurs directly

southeast of the Fault, followed firstly by the Lower Old Red Sandstone shales, sandstones and conglomerates of the Garvock Group, and finally by the sandstones of the Strathmore Group (Macgregor, 1968). Both the River North Esk and River Westwater flow in a gorge through the Fault Zone.

(3) <u>The lower reaches</u> of the rivers lie in the Vale of Strathmore which is composed almost entirely of the Old Red Sandstone series. The Old Red Sandstone largely comprises massive, bright-red, current-bedded sandstones with occasional pebble beds, mottled red sandstones, marly beds and mudstones. The Old Red Sandstone is penetrated by several dolerite dykes and sheets of basalt (hypersthene - and olivine - basalts) and andesitic lavas of the Montrose volcanic centre, mainly located towards the mouth of the River North Esk.

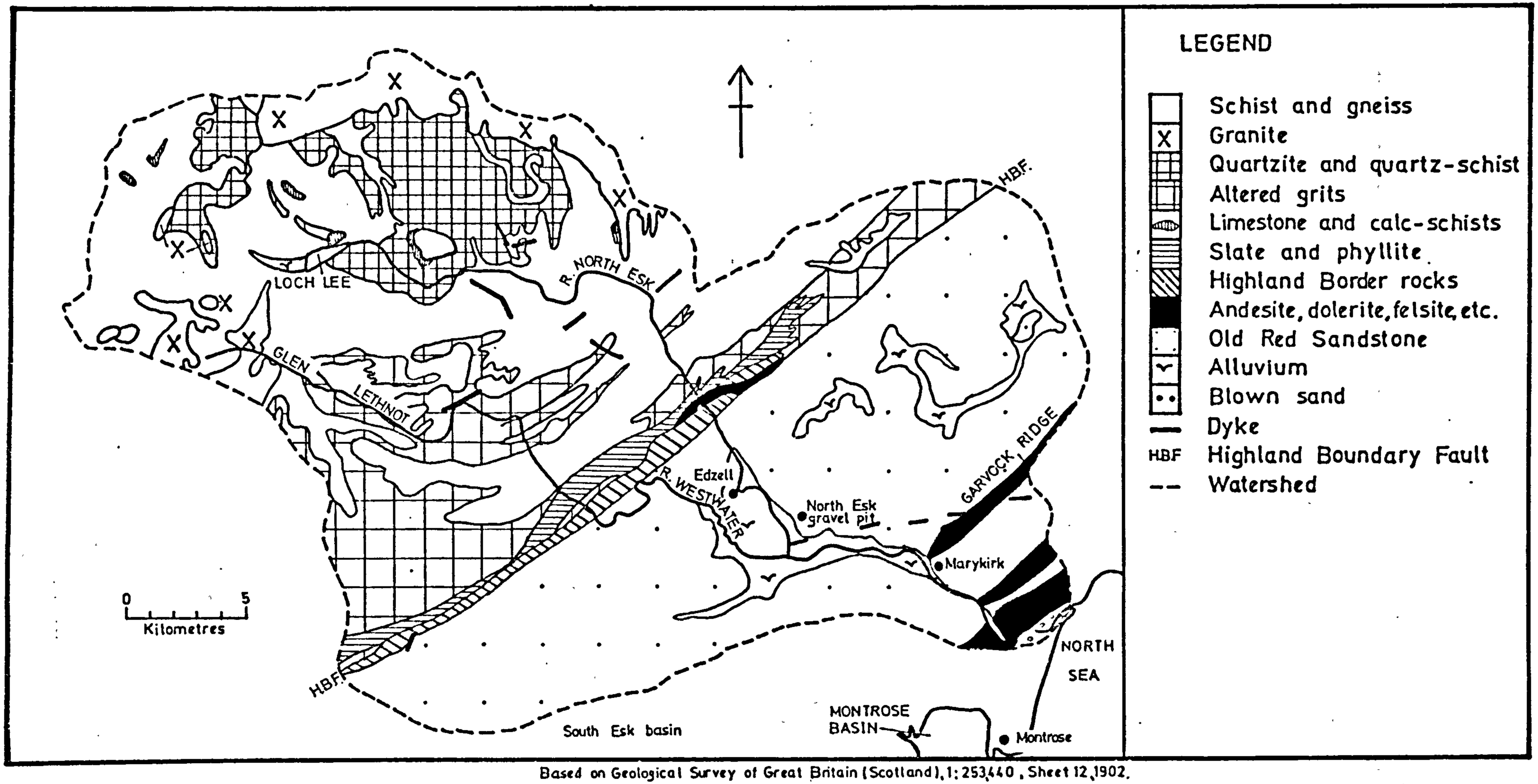
4.2.2.2. <u>Solid Geology of the South Esk basin</u>

The River South Esk also has its source within the intensely folded metasediment region of the crystalline Grampian massif, and traverses the Highland Boundary Fault to flow over the Old Red Sandstone and Devonian lavas of the Vale of Strathmore (figure 4.3b, page 175).

(1) <u>The upper reaches</u> of the South Esk basin cover an area of about 300 km^2. Most of the bedrock lithologies are similar to those of the North Esk and Westwater basins, namely hornblende - and mica-schists, Cairngorm granites, gneisses, altered grits, and smaller outcrops of limestone, felsite and quartz porphyry (Barrow et al, 1912). The 'Green Beds' (Kamp, 1970) and dioritic masses, however, are found only in the South Esk basin.

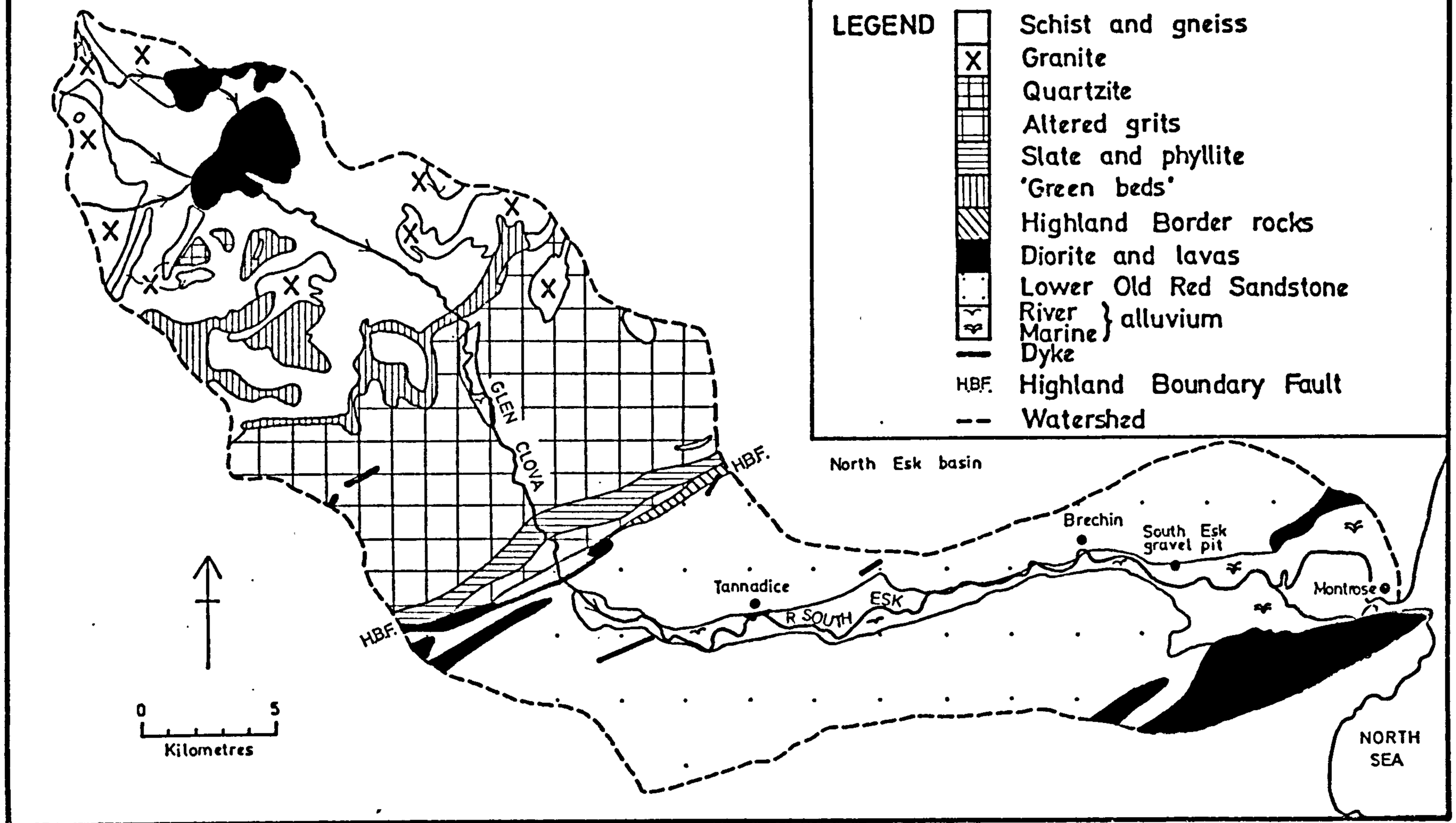
(2) <u>The Highland Boundary Fault Zone</u> extends for some 7 km, and comprises three major lithological zones trending southwest-northeast, namely, the slate and phyllite zone, the zone of altered grits, and finally the Old Red Sandstone which is interrupted by a serpentine dyke.

(3) <u>The lower reaches</u> of the River South Esk lie entirely on Old Red Sandstone bedrock. There are no extents of lava as



Based on Geological Survey of Great Britain (Scotland), 1: 253,440, Sheet 12, 1902.

FIGURE 4.3a SIMPLIFIED SOLID GEOLOGY OF THE NORTH ESK AND WESTWATER RIVER BASINS



Based on Geological Survey of Great Britain (Scotland), 1: 253,440, Sheet 12, 1902

FIGURE 4.3b SIMPLIFIED SOLID GEOLOGY OF THE SOUTH ESK RIVER BASIN

in the North Esk lowland reach, but a basic dyke outcrops near Tannadice. The River South Esk reaches the sea via the Montrose Basin (see Howden, 1870), lying only 6 km south of the mouth of the River North Esk.

4.2.2.3. Solid Geology of Glen Lethnot

The drainage area covered by the waters flowing through Glen Lethnot is only about 31 km^2. The major rock types are schist and gneiss, with minor outcrops of granite and altered grit (figure 4.3a, page 174). Gneiss, granite and altered grit were completely absent from the outwash deposits which were wholly composed of schist and vein quartz (see Chapter 6.3).

4.3. AREAL EXTENT OF THE OUTWASH BODY

4.3.1. INTRODUCTION

The areal extent of a paleo-outwash body is often difficult to determine because the deposits may be represented only by scattered terrace remnants or cannot be easily traced laterally and downstream. Many workers, in particular those carrying out sedimentological studies, have tended to ignore the scale of the deposit while concentrating on a small number of sections (e.g. Illies, 1949; Krumbein and Lieblein, 1956; McDonald and Banerjee, 1971; Martini and Ostler, 1973; Eynon and Walker, 1974). However, it is important that the extent of the outwash deposits, and if possible, the depth too (e.g. Bossons, Chapter 3) should be determined in order to estimate how representative of the whole deposit one's own observations may be. The investigation sites may thence be located in relation to the distance from source, from local 'base-level', and from any geomorphological influences derived from the peripheral regions of the outwash body. The determination of the areal limits of the deposit may also lead to further investigation into the reasons for the existence of those particular boundaries. The deposit may, for example, end gradually or abruptly,

as a result of rising ground or of subsequent fluvial erosion.

4.3.2. THE FOUR SERIES OF SCOTTISH DEPOSITS

The lateral limits of the Strathmore and Glen Lethnot outwash deposits are marked by a break in slope as the surrounding land rises above the outwash level. The downstream limits are more difficult to define. According to Howden (1870) 'high-level gravels' of the North Esk are still found at over 33m OD at about 2 km from the sea (at the Stone of Morphie). The major part of the outwash plain extends only as far as Marykirk, with possible remnants further downstream where Postglacial (Flandrian) river terraces merge with raised shoreline terraces (Crofts, 1974). The River Westwater outwash reaches its downstream limit at its confluence with the River North Esk. The downstream limits of the River South Esk merge with raised shoreline terraces (Cullingford and Smith, 1974) where approximate limits can be defined on the basis of changes in the gradient and sedimentology of the deposits.

The approximate areas covered by each of the outwash bodies under investigation (Table 4.1) give some idea of the magnitude of the differences of scale involved in the comparison of a variety of outwash deposits.

Table 4.1

Areal extent of outwash bodies

Location	Approximate length of outwash body (m)	Approximate areal extent of outwash body (m^2)
1. Bossons	2.5×10^2	10^4
2. Lethnot	2.5×10^3	9×10^5
3. Westwater	7.3×10^3	9.5×10^6
4. North Esk	1.6×10^4	5.5×10^7
5. South Esk	2.6×10^4	6.6×10^7

4.4. THE AGE OF THE DEPOSITS

4.4.1. INTRODUCTION

A major problem in comparison of deposits is the determination of the relative age of each deposit. Differences in dates of formation help to account for variations in the degree of weathering of the sediments, and in any structural deformations which could have occurred (see sections 4.6 and 4.7) over a prolonged period of time. An established date also allows the results or inferences to be incorporated into a regional chronological framework. This suggested sequence of glacial events may itself be subsequently modified, or further supported, by evidence arising from detailed study of the particular deposits.

4.4.2. DEVELOPMENT OF A LATE-WEICHSELIAN CHRONOLOGY

The glacial chronology of the Late-Weichselian, or Devensian, in the Strathmore region of eastern Scotland has not yet been finally established. Early workers such as Agassiz (1840), Buckland (1840) and Lyell (1840) recognised morphological and stratigraphical evidence of former glaciations but Simpson (1933) was the first to identify two glacial readvance stages: the 'Perth Readvance' and the smaller, later 'Loch Lomond Readvance', based on a till-outwash-till stratigraphical sequence. These stages were later attributed to zone I (Lower Dryas) and zone III (Upper Dryas) respectively, of the Late-Glacial pollen sequence, separated by zone II, the Allerød Interstadial (Lacaille, 1950). Charlesworth (1955) also recognised two glacial phases of the Late Glacial, which he named the 'Highland Glaciation', in which ice moved eastwards and extended beyond the east coast, and the 'Moraine Glaciation', possibly correlating with the 'Perth Readvance'. Synge (1956) described similar glacial stages to those proposed by Simpson, and he correlated the extensive outwash gravels of Strathmore with sea-level heights of 33m falling to 8m (100 ft. to 25 ft.) OD during the 'Perth Readvance', and 10m (30 ft.) OD during the 'Loch Lomond', or 'Dinnet', Readvance.

Other workers in the Forth and Tay valleys (e.g. Sissons, 1963,1964; Sissons and Smith, 1965; McLellan, 1969b) reaffirmed the existence of a 'Perth Readvance' glacier and mapped its maximum extent in a number of areas, based on the morphological transition from outwash terraces (beyond the ice-limit) to 'kame-and-kettle' topography (within the ice-limit). The 'Perth Readvance' ice was found to have extended no further than the Forfar moraine in Strathmore, and the mouths of the Grampian valleys. The evidence directly conflicted with evidence presented earlier by Rice (1959b, 1962), who described fluvioglacial features on the Angus coast which he claimed were formed during or after a period of high sea-level (33m). More recently, reinterpretation of stratigraphic sections (e.g. Paterson in the Almond Valley, 1974) and re-examination of morphological evidence (e.g. Francis et al., 1970; Clapperton, 1974) has led to the complete reorganisation of the chronology proposed by Simpson over 40 years ago. It appears likely that zone I times were marked by the presence of an ice-sheet which extended beyond the present east coast of Angus. The ice gradually retreated into the Grampian valleys leaving in its wake a series of extensive outwash sheets intermingled with ice-marginal and subglacial features and small readvance moraines. The so-called 'Perth Readvance' therefore only represents a minor advance during the major ice-sheet recession from the east, which took place between 13000 years BP (Paterson, 1974) and 15000 years BP (Sissons, 1974a).

The major series of outwash deposits studied in the Strathmore region are therefore separated from those forming in the Bossons proglacial valley by at least 13000 years. The North Esk, Westwater and South Esk outwash deposits were formed during the retreat stages of an extensive ice-sheet, by which time ice had become confined in their respective source valleys in the Grampian massif to form 'valley glaciers'.

The existence of the subsequent 'Loch Lomond Readvance' has proved more acceptable to geomorphologists because of the greater freshness of terminal moraines, 'hummocky moraines', meltwater channels and outwash terraces (Simpson, 1933; Sissons, 1973,1974a,b; Sissons and Grant, 1972). Mapping of the probable glacier extents (e.g. Sissons, 1972,1973,1974c; Sugden, 1973) indicates that Grampian glaciers were wholly confined to the heads of the major valleys, and had little effect on the morphology or stratigraphy of the Strathmore outwash series. Dating of this minor readvance is fairly reliable as it is based on carbon 14 dating from several localities, and from pollen analysis from over 60 sites (Sissons, 1967b, 1974b). The outwash deposits, relating to the retreat of the valley glaciers during the 'Loch Lomond Readvance', were formed at least 2000 years later than the Strathmore outwash series, namely, between 10750 and 10250 years BP (Sissons, 1967b; Price, 1974b). Sissons (1972) proposed that the outwash deposits in Glen Lethnot, forming terraces lying several metres above the present river level, date from 'Loch Lomond Readvance' times. It is possible that these high altitude valley-floor deposits (situated above 330m OD) were also modified during the postglacial 'Little Ice Age'. King (1971) dated lichens from solifluction lobes and stripes to between 1730 ($\pm$50 years) and 1830 ($\pm$60 years) A.D. (see also Sugden, 1974). King's data suggest that only two hundred years ago, glaciers, or at least snow and ice, may have been far more extensive and morphologically significant in Scotland than formerly realised.

4.4.3. EVIDENCE OF THE RELATIVE AGE OF THE FOUR SCOTTISH DEPOSITS

Evidence suggesting the relative age of the Scottish deposits investigated in this study is summarised below.

4.4.3.1. The North Esk, South Esk and Westwater outwash deposits

(1) The outwash deposits were found to be underlain by

'red boulder clay' (see also Howden, 1870; Lacaille, 1950; Charlesworth, 1955; W.C. Clarke, personal communication, 1972), suggesting that the 'gravels' represent the outwash formed during the retreat of the last major glacier in Strathmore, i.e. the penultimate glacial phase.

(2) The outwash deposits do not occur below about 27m OD (80 ft. OD), suggesting that the meltwaters were 'graded' to the high sea-level characterising the valley-glacier phase (or 'Perth Readvance') of the ice-sheet recession (e.g. Synge, 1956; Rice, 1959b; Sissons, 1967a). This hypothesis is based on the assumptions that the longitudinal profiles of the outwash deposits were continuous and unbroken (but see Culling, 1957b) and that any lower deposits have not subsequently been removed by fluvial, or even marine, activity.

(3) The outwash deposits are penetrated by fossil ice-wedges and the original bedding is frequently disrupted by cryoturbate features, such as involutions (see later, figures 4.20 and 4.21, page 211). An indurated soil horizon also exists in many areas at a depth of 40-65 cm, probably representing the former permafrost layer (Fitzpatrick, 1956,1958; Common and Galloway, 1958; Galloway, 1961a,b; McManus, 1966; Rolfe, 1966; McLellan, 1969b; Sissons, 1974b; and section 4.7). It has been argued that the periglacial features must have formed in a succeeding cold phase since the outwash deposits were still accumulating during the earlier cold phase, supporting the view that the outwash deposits were formed during the retreat of the penultimate glacial phase (e.g. McLellan, 1969b; Sissons, 1974b; Williams, 1974). However, other evidence indicates (a) that periglacial structures may form syngenetically with the deposition of outwash gravels (e.g. Lavrushin, 1961; Johnsson, 1962; Church, 1970) and do not necessarily represent a later and separate cold phase; (b) that ice-wedge pseudomorphs have not been found on the dissected, lower terraces. Periglaciation, therefore, probably occurred prior to valley entrenchment which began during the

last phases of retreat of the 'penultimate' ice (i.e. zone I, not zone III); and (c) that the succeeding glacial phase, zone III, may not have been long enough (500 years) nor cold enough for permafrost (Williams, 1974), or at least, extensive ground-ice conditions (Pissart, 1970), and associated cryoturbate structures to have developed, although Sissons (1974c) claims that conditions were cold enough for an ice-cap to develop in that short time. The presence of peri-glacial structures alone does not provide conclusive evidence of the date of formation of the outwash deposits, although it does suggest deposition during the penultimate glacial phase.

4.4.3.2. Glen Lethnot

Morphological evidence supporting the view that the outwash deposits of Glen Lethnot were formed during the minor readvance of 'Loch Lomond' times, is summarised below (and see figures 4.2, page 169 , and 4.12, page 194).

(1) The outwash deposits are confined to the head-water valley of the River Westwater (the Water of Saughs) extending over a distance of only 2 to 3 km. Post-depositional dissection by the postglacial river has been on a small scale, the highest terraces lying only 4 to 5m above present river level.

(2) Some of the terraces can be traced into an area of 'hummocky moraines' possibly deposited by stagnating ice. The boundary between these series of mounds and the commencement of the outwash terraces may represent a former ice-margin (Sissons, 1972) (see section 4.5.2.4.).

(3) Marginal fans emerging from lateral valleys on to the terraces appear to be inactive and stable, and are well secured by as dense a cover of low vegetation as on the terraces themselves. The fans may be similar in origin to the 'paraglacial' fans of British Columbia described by Ryder (1971a,b) and by Church and Ryder (1972). These alluvial fans reached maximum development soon after the

disappearance of the last remnants of ice, when sediment supply was greatest, and vegetation development had not yet impeded the redistribution of weathered debris. The fans were probably formed at the end of zone III and during early zone IV, suggesting that the outwash was deposited earlier in zone III. Paraglacial fan aggradation may have taken place prior to, in conjunction with, or subsequent to any fluvial downcutting in the main valley to form the terrace sequences.

4.5. POSITION OF THE ICE-MARGIN

4.5.1. INTRODUCTION

Knowledge of the position of the ice-margin can prove important in several ways. Firstly, it allows one to determine the direction and the distance of fluvial transport from the ice-snout to the site of deposition. Secondly, outwash terraces cut in paleo-deposits which can be traced upstream to terminal moraines can be related to particular glacier advances (e.g. Penck and Brückner, 1909). Thirdly, determination of the whole glacier margin, or at least as much of it as is possible to determine, will provide an estimate of the glaciated area. The extent of the ice-covered area gives a clue as to the relative size of the glacier which has been contributing debris for redistribution by proglacial meltwaters. The glacierized area may also indicate the range of source localities from which the glacially eroded debris is derived, particularly if ice-movements have followed diffluent or transfluent paths (Linton, 1951,1963). Estimates of the distance of transport within the ice could be useful in the interpretation of the size of materials which become part of the proglacial meltwater load as they emerged from the ice-snout (see Chapter 7). Finally, mapping successive positions of the ice-margin provides a record of variations in the length of the glacier (e.g. Matthes, 1935; Field, 1966; Kick, 1966; Petrie and Price, 1966) and of geomorphological changes associated with the glacier recession or series of readvances.

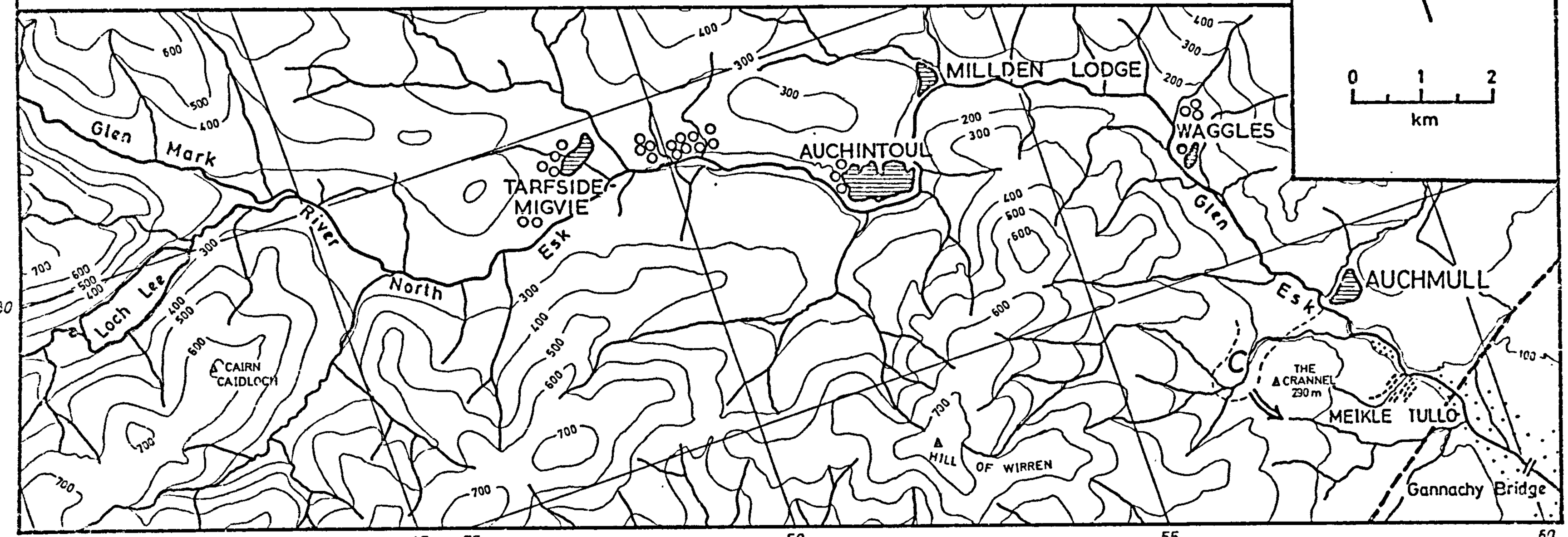
Most modern glaciers can present difficulties in determining the ice-marginal position, particularly if the ice-margin is mantled with rock detritus, or if the marginal moraines consist of high proportions of ice, or if the ice-margin is surrounded by partly detached masses or blocks of ice (e.g. Blachut and Müller, 1966; Field, 1966; Price, 1973). The difficulties can be increased when attempting to define the limits of ancient ice-masses. The position of former ice-margins can only be inferred from morphological and stratigraphical evidence by analogy with present-day features such as the distribution of terminal moraines and 'hummocky moraines', outwash deposits, positions of ice-marginal meltwater channels and termination of subglacial moraine and fluvioglacial phenomena. Problems arise when terminal moraines or outwash deposits are absent, either through erosion, burial, or non-deposition (e.g. Speight, 1940) if the topography is unfavourable (e.g. in a gorge, or on a steep slope) or if meltwaters are sufficiently powerful to remove all new debris brought from upstream.

4.5.2. ICE-TERMINUS POSITIONS OF THE FOUR SCOTTISH GLACIERS

4.5.2.1. The position of the North Esk glacier terminus

The North Esk glacier followed a pattern of oscillating retreat across the Vale of Strathmore, interrupted by minor advances, and periods of stagnation. A series of terminal moraines and areas of 'kame-and-kettle' topography lining Glen Esk provide several probable ice-marginal positions during the formation of the massive outwash deposits (figure 4.4.). These sites include (a) the kame system at Meikle Tullo (figure 4.5, page 186) The ridges are not moraines as suggested by Macgregor (1969) since they are formed of well-stratified sands and gravels (see figure 4.6, page 186). Subglacial waters, probably close to a stagnating ice-margin, flowed south-eastwards towards the glacier axis containing the main meltwater outlets (see also Pitty, 1971, p. 184).

FIGURE 4.4 MAJOR TERMINAL MORAINES IN GLEN ESK



GRID LINES AND BASE-MAP BASED ON ORDNANCE SURVEY SHEET NO. 1: 63360, 1958

(See Sissons (1972) for proposed zone III glacial limits)

- Rivers
- 400 — Contours at intervals of 100 m
- Approximate line of Highland Boundary Fault
- Proglacial outwash deposits associated with ice-limit at Auchmull moraine
- Major terminal moraine in Glen Esk
- Major tracts of hummocky terrain and kame-and-kettle topography
- Meikle Tullo kame system
- Limit of ice-dammed lake ['Lake Crannel'] and associated overspill channel (see Charlesworth, 1955)

Figure 4.5
Part of the kame system near Meikle Tullo, looking towards the west-north-west

Figure 4.6
Crude stratification in the kame deposit near Meikle Tullo

Charlesworth (1955), Synge (1956) and Sissons (1963) consider that the kame system was formed beneath an eastward moving Strathmore glacier. It seems more likely, however, that they simply represent deposition beneath the stagnating ice cover of part of the receding North Esk glacier. There may also have been a plentiful supply of meltwater flowing directly from the overspill draining ice-dammed 'Lake Crannel' (Charlesworth, 1955), which subsequently flowed into and beneath the ice.
(b) The site selected as representing the ice-snout of the North Esk glacier during formation of much of the outwash body is that of the large terminal moraine close to the settlement of Auchmull. The Auchmull moraine lies about 800m upstream of the first preserved outwash terrace remnant (Sissons, personal communication, 1972).
(c) Charlesworth (1955), by contrast, claims that the North Esk glacier lay 'well within its valley', namely at the Millden Lodge Moraine which lies about 8 km upstream from the Auchmull moraine. His argument is based on the abundance of moraines of the 'Moraine Glaciation' (Stage 1 of his glacial sequence) above Millden and the absence of moraines below it.

The moraines of Glen Esk clearly represent a general, if oscillating, glacier retreat upstream from Meikle Tullo, with major halt stages at Auchmull and Millden Lodge, and minor halts at other sites such as Auchintoul (Macgregor, 1968) and Migvie (Synge, 1956) (figure 4.4.).

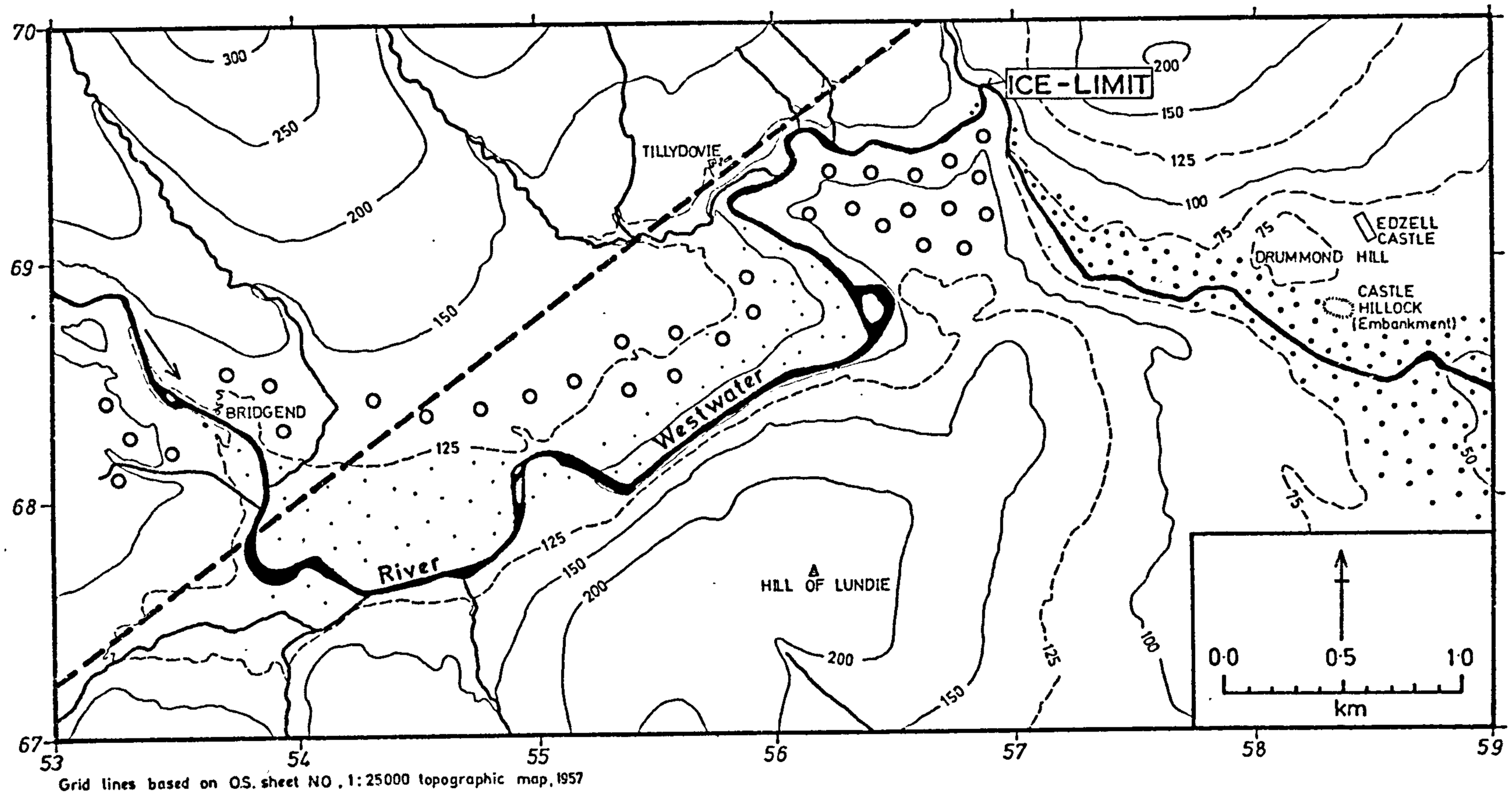
Sissons (1972) has plotted the probable ice-margins of the cirque glaciers of the 'Loch Lomond' glacial re-advance, and the approximate sites of related outwash terraces. None of these zone III features extend within 3 km of even the Migvie Moraine. The North Esk outwash deposits are clearly not contemporaneous with the zone III deposits, nor have they been modified by subsequent meltwaters from zone III glaciers.

4.5.2.2. The position of the Westwater glacier terminus

The valley floor of the River Westwater is marked by a series of outwash terraces, separated from each other laterally by terrace bluffs running parallel with the present river and longitudinally by a series of valley constrictions in which no deposition occurred. Field work in the valley of the River Westwater shows that it contains several important recessional moraines and numerous examples of subglacial moraine, and 'kame-and-kettle' features, reflecting a complex history of deglaciation (figure 4.7, page 189). A conspicuous kame-moraine system occurs at Bridgend where the narrow mountain valley emerges on to the first major outwash terrace of the Westwater. A narrow gorge constricts flow and has prevented any outwash deposition beyond Tillydovie where the river flows along the Highland Boundary Fault Zone.

The second major series of outwash terraces radiate from the end of the valley constriction, while morainic debris plasters the gentle upper slopes of the overlooking valley sides (figure 4.8, page 190). 'Kame-and-kettle' topography overlooking outwash terraces (figure 4.9, page 191), was also observed along the edge of the Highlands around Edzell Castle, marking the terminal zone of the Westwater glacier at one stage in its recession. The ice-snout of the Westwater glacier probably receded from 'Castle Hillock' upstream, with its halt-stages marked by the series of moraines extending far upstream of the Bridgend system. The ice-snout position during formation of the Westwater outwash deposits was therefore variable. The ice-snout position selected for the purpose of later calculations lay at the mouth of the valley constriction. The second major outwash series which commences at this point is contiguous farther downstream with the outwash deposits of the River North Esk (see figure 4.10).

FIGURE 4.7 MORPHOLOGY OF THE LOWER WESTWATER VALLEY



Grid lines based on O.S. sheet NO. 1:25000 topographic map, 1957

Rivers
100 Contours at intervals of 50 m
25 Contours at intervals of 25 m
Approximate line of Highland Boundary Fault
Major tracts of hummocky terrain and kame-and-kettle topography
Proglacial outwash deposits associated with ice-limit as marked
Proglacial outwash deposits associated with ice-limit at Bridgend

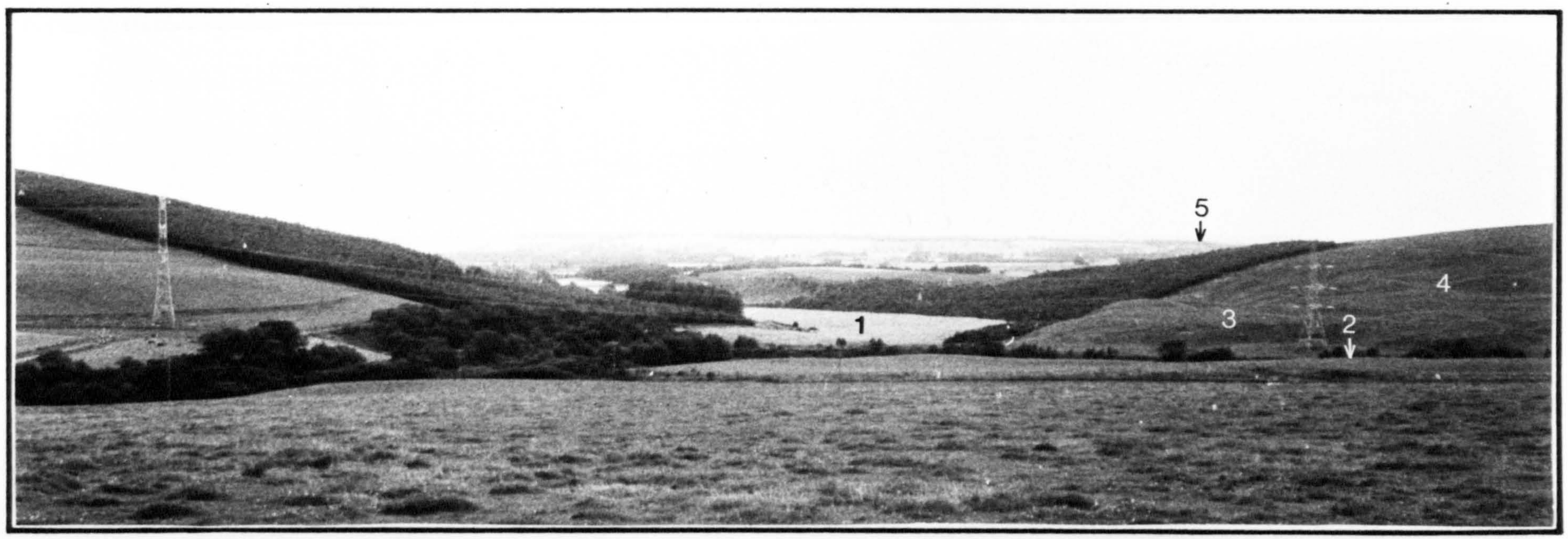
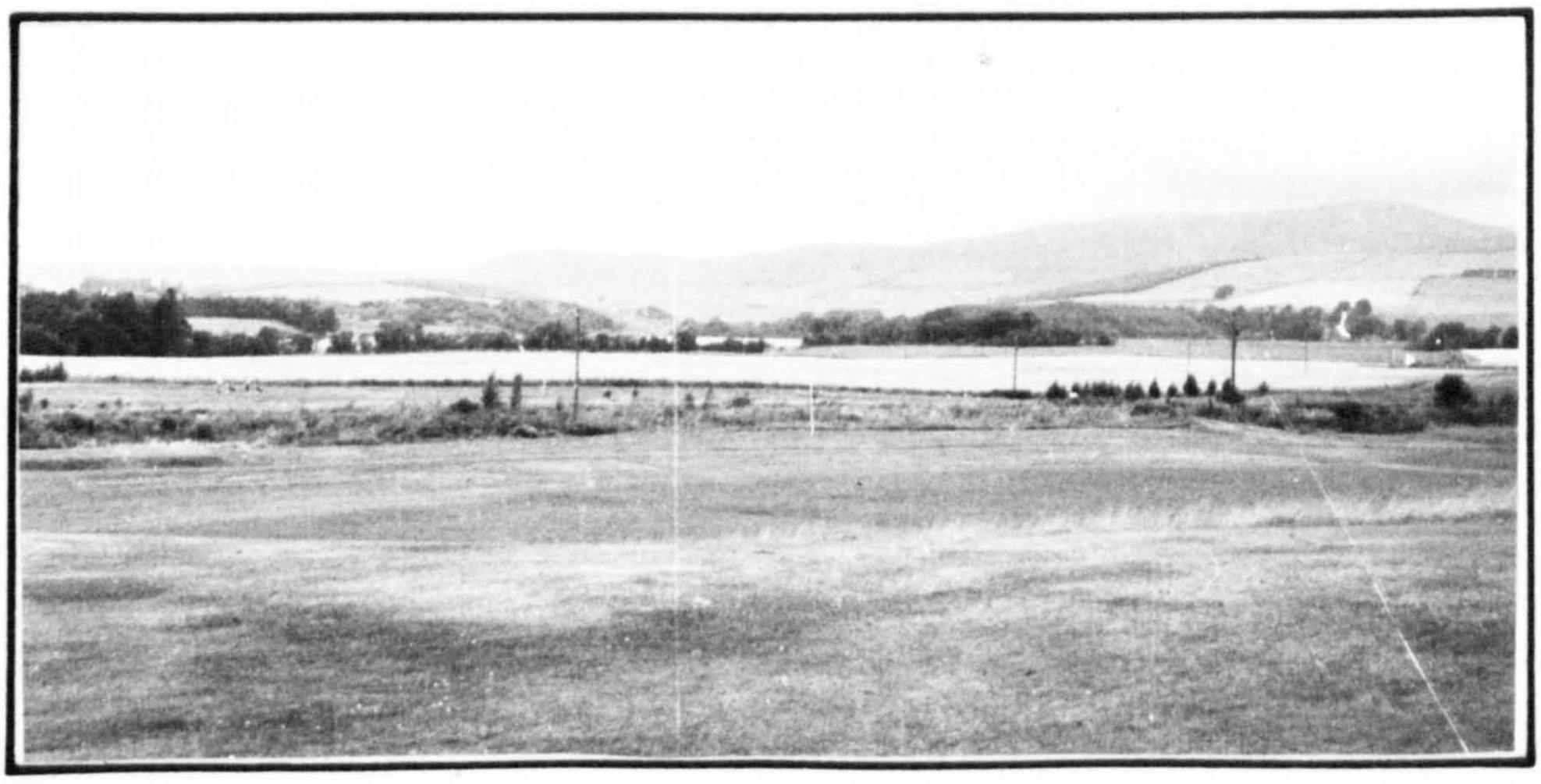


FIGURE 4.8 THE LOWER OUTWASH TERRACE IN THE WESTWATER VALLEY

1 Outwash terrace, looking downstream; 2 River Westwater; 3 Former ice-marginal zone; 4 Hummocky terrain; 5 River North Esk valley

FIGURE 4.9
OUTWASH TERRACES 1·5 KM FROM THE WESTWATER ICE-LIMIT

FIGURE 4.10
WESTWATER OUTWASH TERRACES,
LOOKING UPSTREAM FROM EDZELL GOLF COURSE

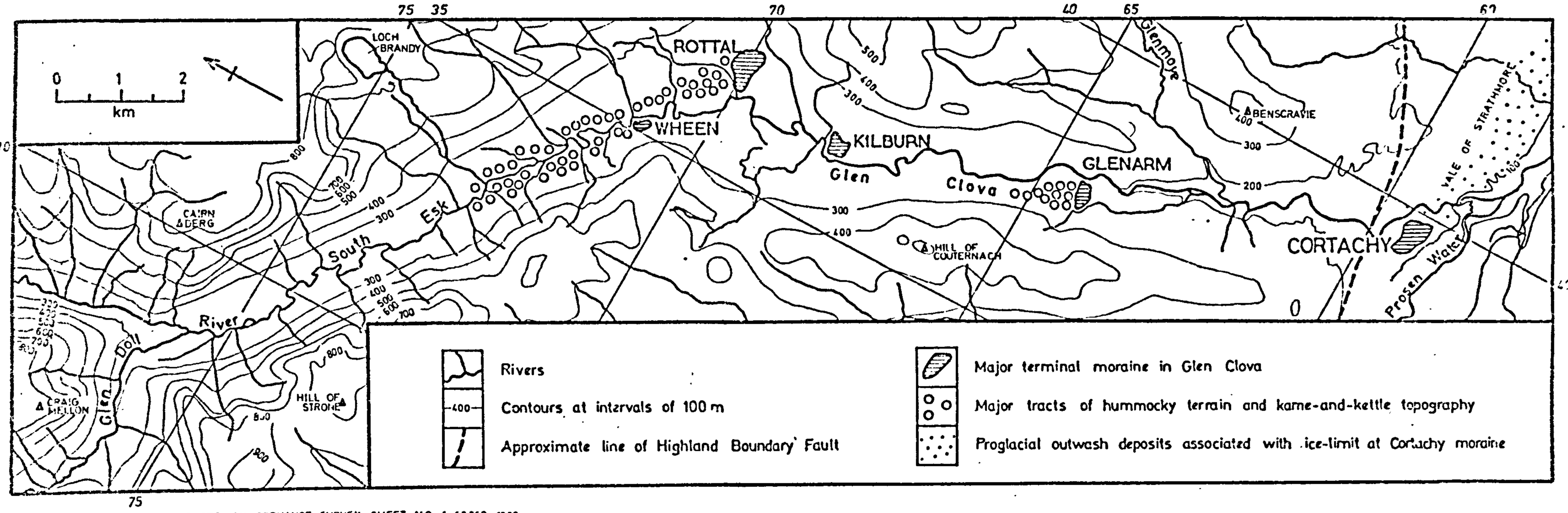


4.5.2.3. The position of the South Esk Glacier terminus

Many retreat moraines occur within the valley of the River South Esk and different workers have claimed different moraines as marking the limits of the glacier. Both Charlesworth (1955) and Synge (1956) regard the moraine at Rottal as the downstream limit of the South Esk glacier during this advance (figure 4.11). Further conspicuous moraines exist downstream, however, including the 'Glenarm Moraine' and the 'Cortachy Moraine'. Each stage in the progressive retreat of this glacier, as in the case of the other glaciers discussed, across the Vale of Strathmore, was associated with a further series of subglacial deposits, moraines and outwash deposits. The final stages of retreat in each case was also accompanied by subsequent stream degradation into the fluvioglacial deposits (see section 4.6). The outwash deposits of the River South Esk have been related in this instance to an ice-margin lying at the mouth of its Grampian valley (Glen Clova), at the 'Cortachy Moraine'.

4.5.2.4. The position of the glacier terminus in Glen Lethnot

The limit of the valley glacier occupying Glen Lethnot during 'Loch Lomond' times is marked, according to Sissons (1972) by the abrupt transition from 'hummocky moraines' to a low outwash terrace. However, definition of this limit in the field is complicated by the concentration of 'hummocks' on the lower valley-side slopes rather than on the valley floor (figures 4.2 and 4.12), by the indeterminate composition, structure and origin of the 'hummocks' (although probably deposited from a stagnating ice mass), by the presence of composite series of terraces rather than of a single terrace, and by the 'contamination' of the peripheries of the valley floor by 'paraglacial' alluvial fans. A possible remnant of a former terminal moraine, situated on a continuous terrace feature, lies opposite the confluence of the Water of Saughs with a stream derived from a 'hanging valley'. The lip of the hanging valley is formed of a conspicuous



GRID LINES AND BASE-MAP BASED ON ORDNANCE SURVEY SHEET NO. 1:63360, 1969

FIGURE 4.11 MAJOR TERMINAL MORAINES AND HUMMOCKY TERRAINS IN GLEN CLOVA

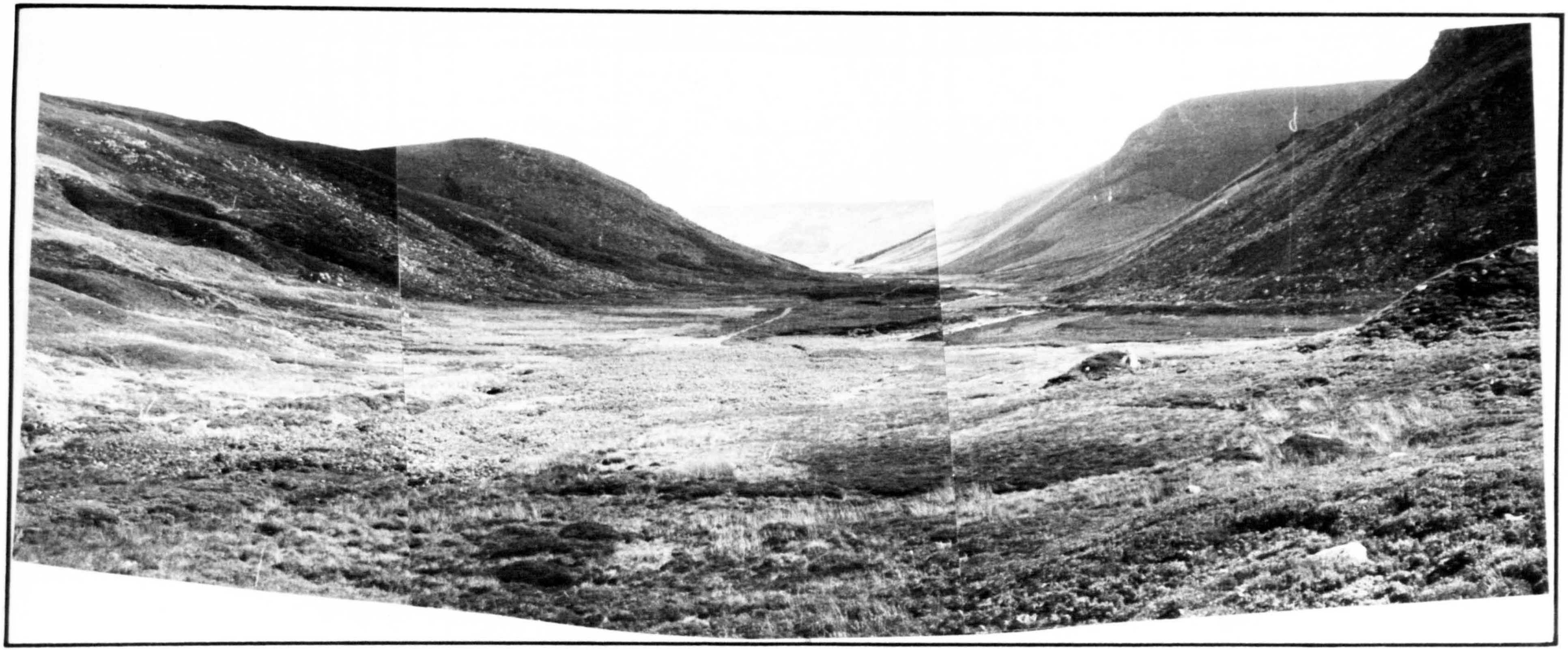
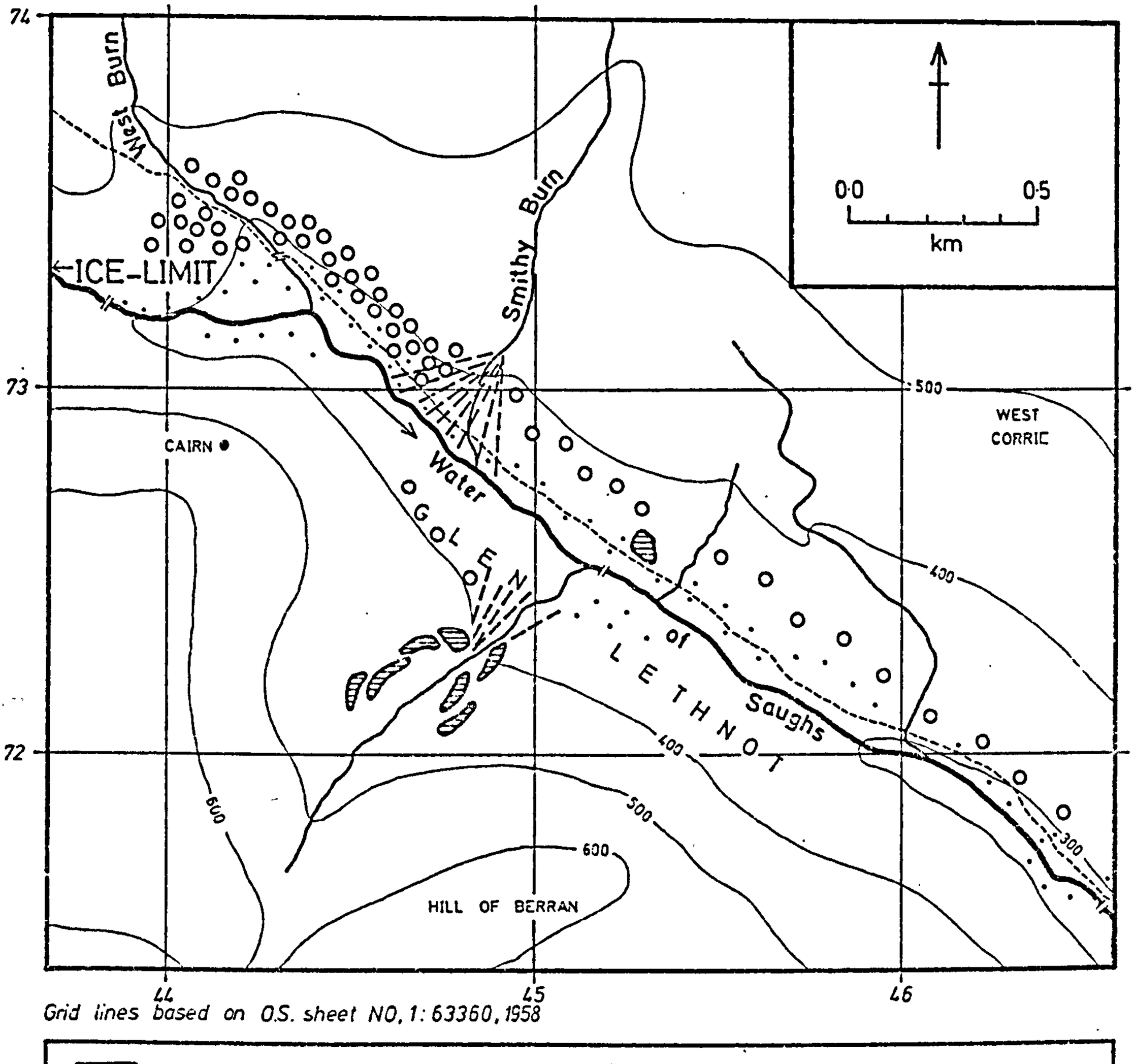


FIGURE 4.12 'HUMMOCKY MORAINE', VALLEY-SIDE FANS AND OUTWASH TERRACES IN GLEN LETHNOT

FIGURE 4.13 SIMPLIFIED MORPHOLOGY OF GLEN LETHNOT



Grid lines based on O.S. sheet NO, 1: 63360, 1958

Streams
300 Contours at intervals of 100 m
Track
Footbridge
Major terminal moraine
Major tracts of hummocky terrain
Outwash deposits
Valley-side fan

terminal moraine (Sissons, 1972). The Glen Lethnot ice terminal has been regarded as lying at the most upstream point where terraces have formed, at an altitude of about 410m (figure 4.13, page 195).

4.6. POST-DEPOSITIONAL MODIFICATIONS: I. ALTERATIONS TO THE LONGITUDINAL PROFILE

4.6.1. INTRODUCTION

Measurement of the gradient of an outwash surface can have important implications for the interpretation of morphological and hydrological features. The shape of the longitudinal profile, for example, reflects the balance between the degree of aggradation and entrenchment that has taken place. The profile partially results, therefore, from the downstream change in the discharge of water, in the quantity and nature of the debris supply, and in the relation between discharge and stream load (e.g. Yatsu, 1955; Fahnestock, 1963; Meier, 1965; Davies, 1969). The size of the particles forming the bedload is closely related to channel slope (e.g. Shulits, 1941; Yatsu, 1955; Hack, 1957; Brush, 1961; Culling, 1957a), which in turn often closely approximates the gradient of the outwash surface (e.g. Church, 1970; Boothroyd, 1972). Measurements of gradient can, therefore, also help in explanations of downstream changes in particle size (Chapter 7).

The nature of the load, discharge and gradient also influence the channel pattern, i.e. whether it is meandering, straight or braided (Leopold and Wolman, 1957; and see Chapter 5), so that knowledge of the paleo-slope can provide further information on stream characteristics. Measurement of the local slope also allows calculation of the true dip of primary sedimentary structures such as cross-bedding or cross-laminations (e.g. Jopling, 1966). The gradient of the local surface is also required for the calculation of total discharge values, including estimates of paleo-discharges, which are partially dependent on local slope values (e.g. Jopling, 1969; Church, 1970).

Finally, measurements of the longitudinal profile can indicate any major slope junctions between stream reaches. A major change in gradient may result from changes in the rates of deposition or degradation as the channel responds to different hydrodynamic conditions (e.g. Culling, 1957b; Krigstrom, 1962; Fahnestock, 1969; McDonald and Banerjee, 1971). The discontinuities in the long profile may reflect (a) a change from the zone of direct ice-marginal influence, i.e. the 'proximal zone' (Chapter 3), where rates of fluvial sedimentation are largely controlled by glacial and climatic regimes, to a zone either of predominantly fluvioglacial activity, i.e. the 'distal zone', or of base-level control (Clayton, 1975). In the latter case, the break of slope represents a 'knick-point' which is gradually migrating upstream; and (or) (b) geological controls, such as a resistant rock bar, in which case the rate of degradation is retarded compared with neighbouring reaches.

The gradient of the topographic surface can be directly surveyed in an area of present-day outwash, and longitudinal profiles of different parts of the outwash, or of the outwash at successive periods can be derived quite simply from the altitudinal data obtained from surveys (e.g. Church, 1972; Boothroyd, 1972; Fahnestock, 1969, and Chapter 3). The measurements represent the altitude at which fluvioglacial sediments have been deposited on the surface, usually following flood discharges and overbank flow of meltwaters (discussed in greater detail in Chapter 5). The resultant aggradational surface can be traced upstream to the ice-margin or to ice-marginal landforms such as terminal moraines, debris fans or 'kame-and-kettle' topography (e.g. Sharp, 1960; Suggate, 1963; Seret, 1965; Sissons and Smith, 1965; Bogacki, 1973). The surface represents the active outwash tract of a glacier whose terminus is situated at a given locality upstream.

The geomorphological situation is substantially modified when considering paleo-deposits, for they have frequently undergone major post-depositional modifications to their original surface profile through two mechanisms; terrace formation and isostatic readjustment. The problems of interpretation of the outwash gradient associated with the effects of terrace formation and of isostatic readjustment are discussed in greater detail in relation to the four Scottish deposits which were investigated.

4.6.2. TERRACE FORMATION AND OUTWASH PROFILES

4.6.2.1. Terrace formation

There has been much discussion in the literature about the origin and significance of both terrace formation and terrace profiles. The highest terrace is generally considered to represent the last surface to be occupied by aggrading meltwater streams. Lower terraces appear to result from periodic downcutting by streams into the previously aggraded surface.

The profile of the highest terrace represents the slope formed by the aggrading proglacial meltwater streams, while that of the lower terrace represents the slope produced by degrading streams, eroding and reworking earlier proglacial deposits. The extent to which the lower terrace surface is composed (a) of undisturbed material derived from the original period (or periods) of aggradation, (b) of reworked material from the underlying deposits, and (c) of new contributions of material from upstream, remains an uninvestigated problem. Greater consideration has been given in the literature to the probable causes of stream entrenchment. The change in stream process from one of predominant aggradation to one of channel erosion results from a change in the relation between discharge and sediment load of the stream. The stream adjusts to new hydraulic conditions by changing its channel slope. Valley dissection results from an increase in stream energy available for vertical erosion either during (a) an increase in the ratio of the volume

of sediment to the volume and competence of meltwater, through a decrease in the debris supply (Church, 1970) or an increase in meltwater discharge (Williams and Rust, 1969) and/or during (b) a fall in base-level relative to land-level (Cotton, 1942; Gage, 1958; Church and Ryder, 1972). The state of the glacier budget, itself dependent on climatic conditions, will largely determine the quantity of meltwater and debris discharged into the proglacial area. It is not known whether or not a retreating glacier gives rise to smaller quantities of meltwater and sediment than an actively eroding, advancing glacier (see Kasser, 1973), but many workers consider that widespread proglacial terrace formation occurred during periods of ice recession (e.g. Cotton, 1942; Hoppe, 1959; Kirby, 1961; John and Sugden, 1962; Seret, 1965,1968; Vivian, 1967a; Ehrlich and Davies, 1968; Slatt and Hoskin, 1968; King and Buckley, 1969; Williams and Rust, 1969; Church, 1970; Jewtuchowicz, 1971; Church and Ryder, 1972). An ice retreat would, therefore, be associated with (a) a reduction in sediment supply and a proportionately smaller reduction in meltwater discharges as the glaciated area progressively decreases (e.g. Church and Ryder, 1972), or (b) an increase in sediment supply and a proportionately greater increase in meltwater discharge, as extensive areas, rich in unconsolidated debris, are exposed by a wasting ice-mass (e.g. Williams and Rust, 1969).

The whole pattern of stream erosion and deposition may be altered if the change in the discharge-sediment load relation is sufficient to replace the braided stream network by a single, sinuous meandering stream (Denton and Stuiver, 1967; Schumm, 1968; Gigout, 1969). Stream discharge may also be increased relative to sediment supply by a steepening of the overall gradient, following a relative lowering of base-level. The influence of a fall in base-level will first be expressed in downstream reaches, gradually extending upstream to reaches in which meltwater and sediment conditions are largely governed by the glacial budget.

Terrace formation thus results from a regrading of the channel profile as a response by streams flowing under new hydraulic conditions. Conditions favourable for dissection of outwash deposits may be produced by a combination of

(a) climatic changes (effecting, for example, glacier retreat) with consequent excess stream energy after sediment transport has been completed;

(b) structural changes in the relative levels of land and sea, which result in increases in gradient and the length of the stream course; and

(c) 'random' changes in discharge and sediment supply. Gage (1970), for example, demonstrated that terraces may be excavated very rapidly (e.g. over a period of 30 years) as a result of 'infrequent high-energy episodes', representing local proglacial hydraulic changes after a major flood.

4.6.2.2. Longitudinal profiles of the Scottish outwash terraces

The outwash deposits of the four Scottish areas investigated have been dissected to form terrace surfaces (see figures 4.14 and 4.15, page 201). Cullingford and Smith (Smith, personal communication, 1973; Cullingford and Smith, 1974) are nearing completion of the levelling of the North and South Esk outwash deposits, which may produce a series of sub-parallel long profiles. A single long profile has, however, been adopted in the present study to represent each major outwash surface in which the series of observation and sediment sample sites are located. The need for accurate levelling was not regarded as vital at this stage as a result of the following procedure:

(1) Sample sites were located on the same terrace surface as far as could be ascertained from field observations, aerial photographs and large-scale published maps (mainly 1:10560);

(2) The height of each site was calculated from large-scale maps;

FIGURE 4.14
OUTWASH TERRACES OF THE RIVER NORTH ESK, LOOKING NORTHWARDS FROM THE NORTH ESK GRAVEL PIT

FIGURE 4.15
OUTWASH TERRACE OVERLOOKING THE PRESENT FLOOD-PLAIN TERRACE OF THE RIVER NORTH ESK

(3) The height of each site was then plotted graphically to allow calculation of the longitudinal profile formed by the sequence of terrace sites, based on regression analysis (e.g. Kirby, 1964a; Sissons and Smith, 1965; Church, 1972) (figure 4.16). The sediment samples from each site can therefore be directly related to up- and down-stream sites on the same profile, assuming (a) that the profile is the correct one and (b) that errors of measurement are not significant. The scale on which the profile equations are calculated is sufficiently small that errors of even one metre will hardly affect the equation. The range of possible profiles is in any case quite small because the outwash deposits (a) are overlain by up to 0.5m of 'topsoil', and (b) do not exceed a maximum of 6 - 7m in thickness[1] (Howden, 1870; Macgregor, 1968; Clarke, 1972, personal communication (Gravel Pit excavators); author's own measurements). The profiles (figure 4.16) are therefore considered to be sufficiently accurate for the present study.

All the profiles exhibit an exponential decrease in altitude downstream (figure 4.16). Two major slope segments are present in all the outwash surfaces, and the difference in slope between the 'proximal' and 'distal' outwash surfaces is significant in each area (Table 4.2).

The 'proximal' slopes extend 0.25 to 2.5 km beyond the respective postulated ice-marginal positions, and are 4.13 to 8.25 times steeper than the 'distal' slopes. No significant change of gradient occurs in the North Esk outwash deposits downstream of their former confluence with the Westwater outwash deposits (figure 4.16).

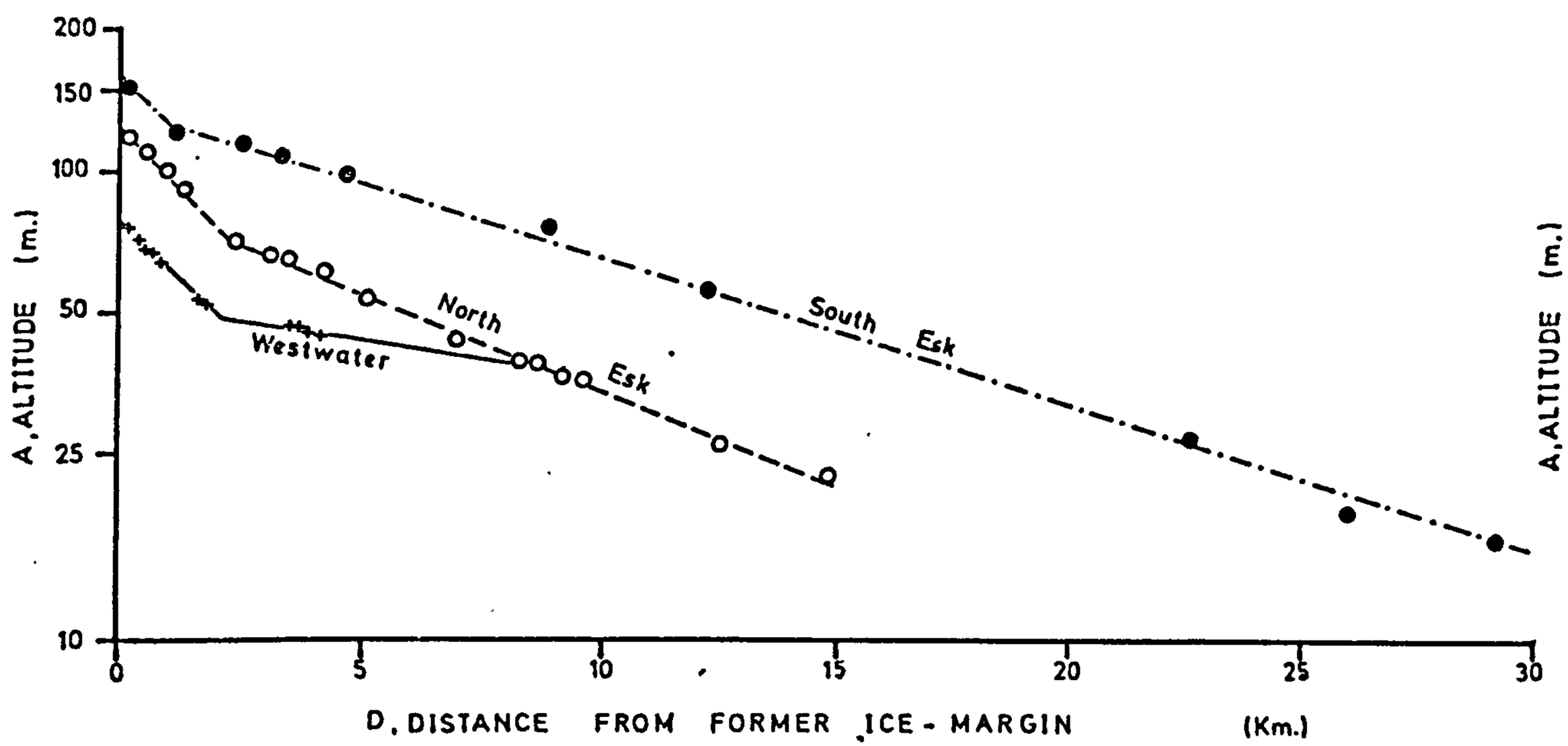
---

1. The North Esk and Westwater outwash deposits are generally less than 3m thick; 2.83m of outwash sediment was exposed in a pipeline excavation situated 1.05 km south of the North Esk gravel pit (GR 640665).

Figure 4.16
Longitudinal profiles of the four Scottish outwash deposits

		PROXIMAL SLOPE	DISTAL SLOPE
—·●—·	South Esk	$A = 153{\cdot}1\ D^{-0{\cdot}10}$	$A = 138{\cdot}2 D^{-0{\cdot}03}$
—○—	North Esk	$A = 125{\cdot}0\ D^{-0{\cdot}10}$	$A = 87{\cdot}7 D^{-0{\cdot}04}$
—+—	Westwater	$A = 75{\cdot}3\ D^{-0{\cdot}11}$	$A = 53{\cdot}5 D^{-0{\cdot}02}$

	PROXIMAL SLOPE	DISTAL SLOPE
Glen Lethnot	$A = 41{\cdot}2\ D^{-0{\cdot}13}$	$A = 38{\cdot}6\ D^{-0{\cdot}04}$



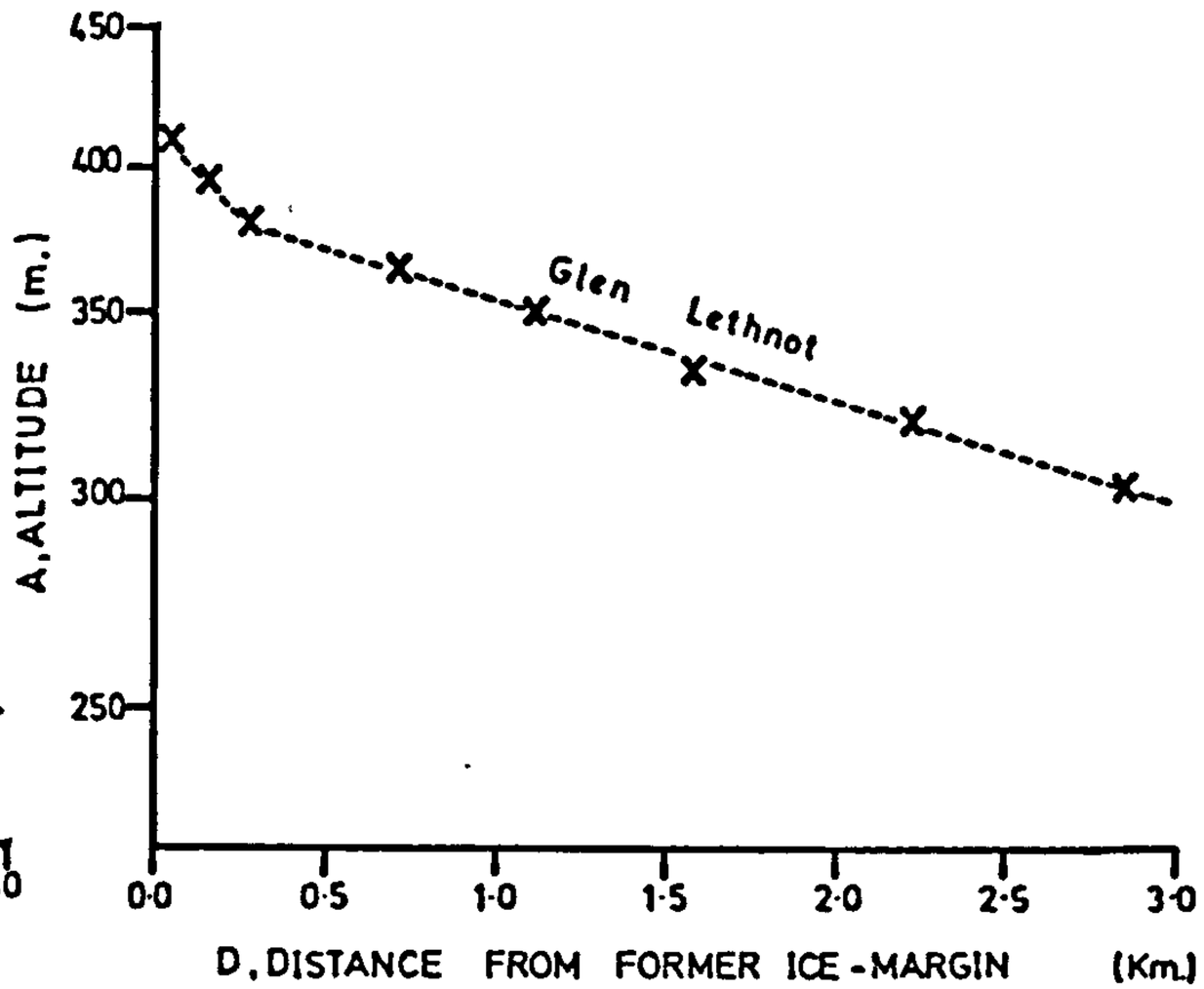


Table 4.2

Average gradient of outwash surfaces

Outwash surface	'Proximal' slope (m/m)	'Distal' slope (m/m)
North Esk	-0.0231	-0.0037
Westwater	-0.0140	-0.0023
South Esk	-0.0305	-0.0037
Glen Lethnot	-0.1240	-0.0300

4.6.3. ISOSTATIC READJUSTMENT AND OUTWASH PROFILES

A relative fall in base-level since deposition of the outwash deposits has resulted from a rise of both land and sea levels, in which the rate of land uplift has exceeded that of sea-level rise (Godwin, 1963). Many Pleistocene proglacial outwash deposits have been tilted as a result of differential isostatic uplift of land, but few workers have attempted to determine the resultant change in gradient of uplifted outwash terraces (see Schumm, 1971). The main reason for the lack of estimates of the increase in gradient since deposition is that few reference levels are available from which to determine the original gradient. Some workers (e.g. Speight, 1963) have calculated the amount of uplift from the profile of the present stream. The assumption that the present and former profiles were a similar shape is, however, not acceptable, for it is likely that the former streams had higher discharges, greater sediment loads, braided patterns and also steeper gradients than the present streams (Culling, 1957a,b; Schumm, 1971). The extrapolation of ancient profiles to determine the former base-levels is also an untenable method, although it has been applied in principle by many other workers simply by assuming parallelism with the present profile.

Many workers have accurately measured the amount of isostatic uplift in coastal areas by the levelling of raised shoreline terraces (e.g. see discussions of literature in Stephens and Synge, 1966; West, 1968). More recently, Andrews (1968,1969) calculated a series of isobases for most of arctic Canada, over the Postglacial period of almost 10,000 years. Church and Ryder (1972) estimated from Andrews' data that an actively aggrading terrace surface in Baffin Island (which reached the head of Ekalugad Fjord) was uplifted 23m over a 1400-year period of Postglacial time (5700 to 4300 years BP). However, outwash deposits which extend some distance inland towards the centre of isostatic uplift will have undergone differential rising, with maximum increases in altitude furthest from the present coast. The influence of features such as the Highland Boundary Fault line in Scotland on rates of uplift is, however, unknown (see Walton, 1966) but is likely to produce irregularities in the terrace profile (see Sissons, 1971). The degree of uplift should be indicated at different points along the whole profile rather than citing a single value, unless it can be demonstrated that differential uplift has not been important over the area concerned.

The raised shoreline terraces of the east coast of Scotland, especially the Firths of Forth and Tay, have attracted considerable attention in recent years. Isobases of the most distinctive shorelines, the Main Perth and the Main Postglacial shorelines, have been constructed in the Forth and Tay regions on the basis of precise levelling of large numbers of shoreline features (e.g. Sissons and Smith, 1965; Cullingford and Smith, 1966; Sissons, 1962,1966, 1967a,1971; Sissons, Cullingford and Smith, 1966; Cullingford and Smith, 1974). The Main Perth shoreline has been shown by these authors to merge upstream, in some river valleys, with extensive outwash terraces, including that of the South Esk (see also Crofts, 1974). The extrapolation of isobases for the Main Perth shoreline inland across the North and South Esk outwash area, which are equivalent in age,

may therefore, indicate the degree of uplift since deposition at different distances from the present coastline. The former gradients of the Scottish deposits investigated have been estimated by extrapolation of isobases of the Main Perth shoreline as published by Sissons (1967a and by Smith et al. (1969) for East Fife (figure 4.17). More recent information presented by Cullingford

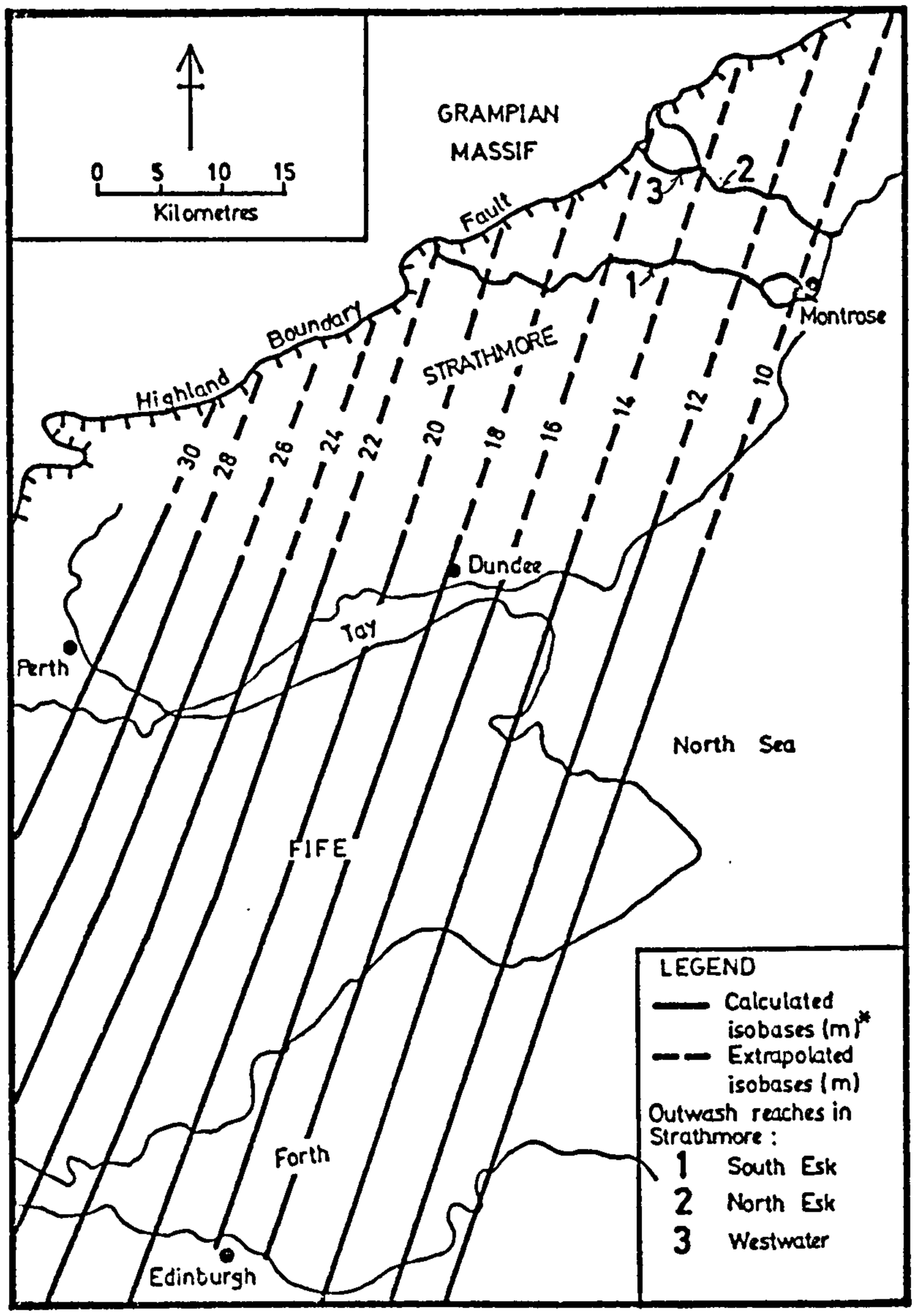


Figure 4.17 Extrapolation of isobases for the Main Perth Raised Shoreline across three Strathmore outwash deposits

*After Sissons, 1967a ; Smith et al, 1969

and Smith (1974, unpublished) suggests that the isobases follow a north-south trend more closely than the north-northeast-southsouthwest trend portrayed in figure 4.17, with the 20m isobase situated at the North Esk estuary. The calculations of changes of gradient (figure 4.18) illustrate a method of determining post-depositional increases in outwash gradient, and in this case they represent the minimum increases in gradient resulting from differential uplift of land relative to sea-level since outwash deposition associated with the 'Main Perth Raised Shoreline'.

The gradients of the original outwash surfaces (figure 4.19 and Table 4.3) can be compared with their present-day profile equivalents (see figure 4.16, page 203 and Table 4.2, page 204). Vertical displacement has not resulted in significantly steeper gradients, and the present outwash surfaces follow very closely the original exponential profiles (figure 4.19, and Table 4.3).

Table 4.3

Minimum changes in gradient resulting from isostatic uplift since deposition of the North Esk, Westwater and South Esk outwash deposits, and formation of the 'Main Perth Raised Shoreline'

Location	Gradient of former surface (m/m)		Change in gradient			
			Actual change (m/km)		Percentage change	
	'proximal'	'distal'	'proximal'	'distal'	'proximal'	'distal'
North Esk	-0.0228	-0.0033	+0.360	+0.383	+ 1.58	+ 11.45
Westwater	-0.0136	-0.0019	+0.340	+0.380	+ 2.50	+ 19.94
South Esk	-0.0301	-0.0034	+0.400	+0.357	+ 1.33	+ 10.57

Differential isostatic uplift has therefore not proved to be significant in altering the actual gradient of the former outwash surfaces, for the whole surfaces have been uplifted to

Figure 4.18
Post-depositional uplift of the Scottish outwash deposits

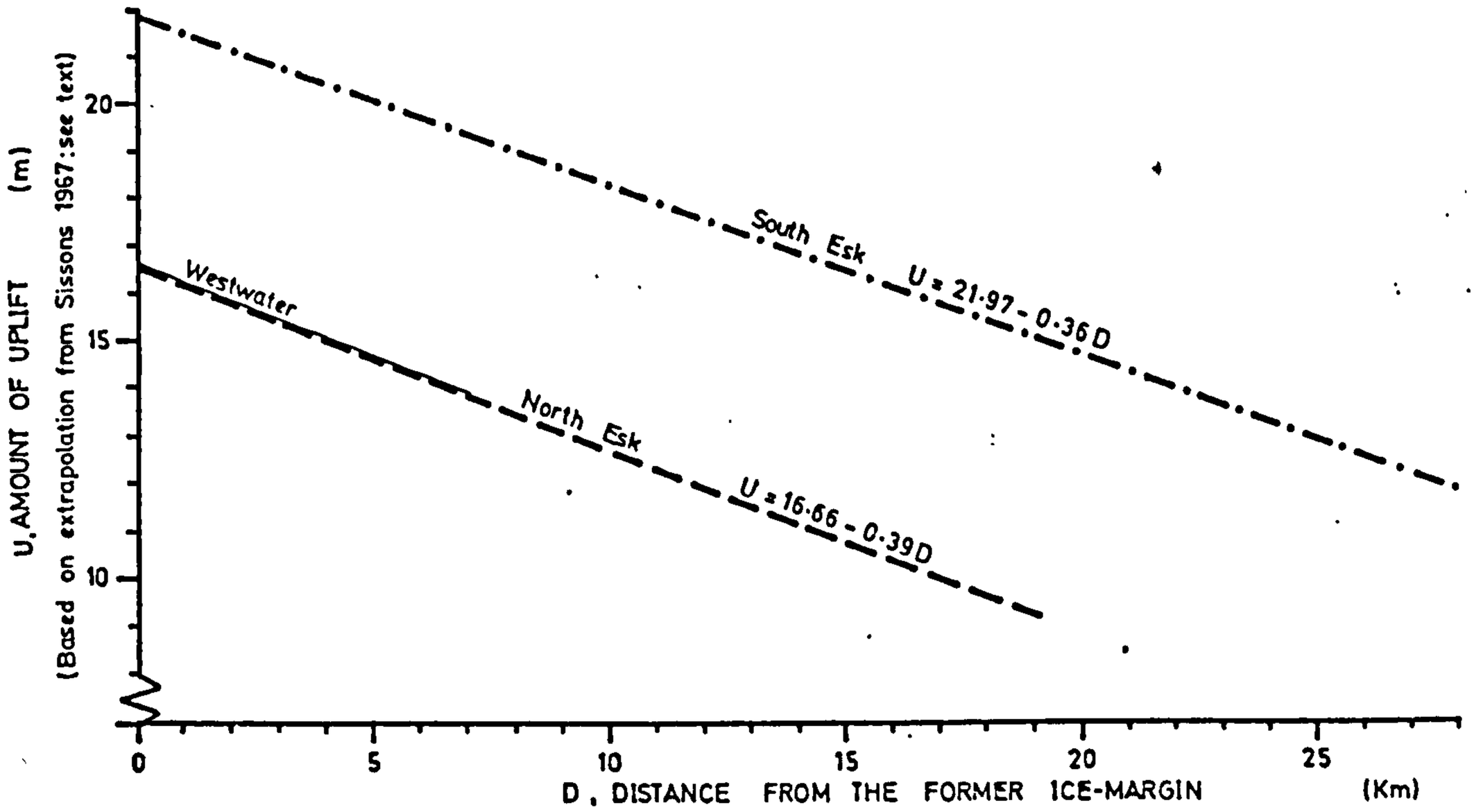
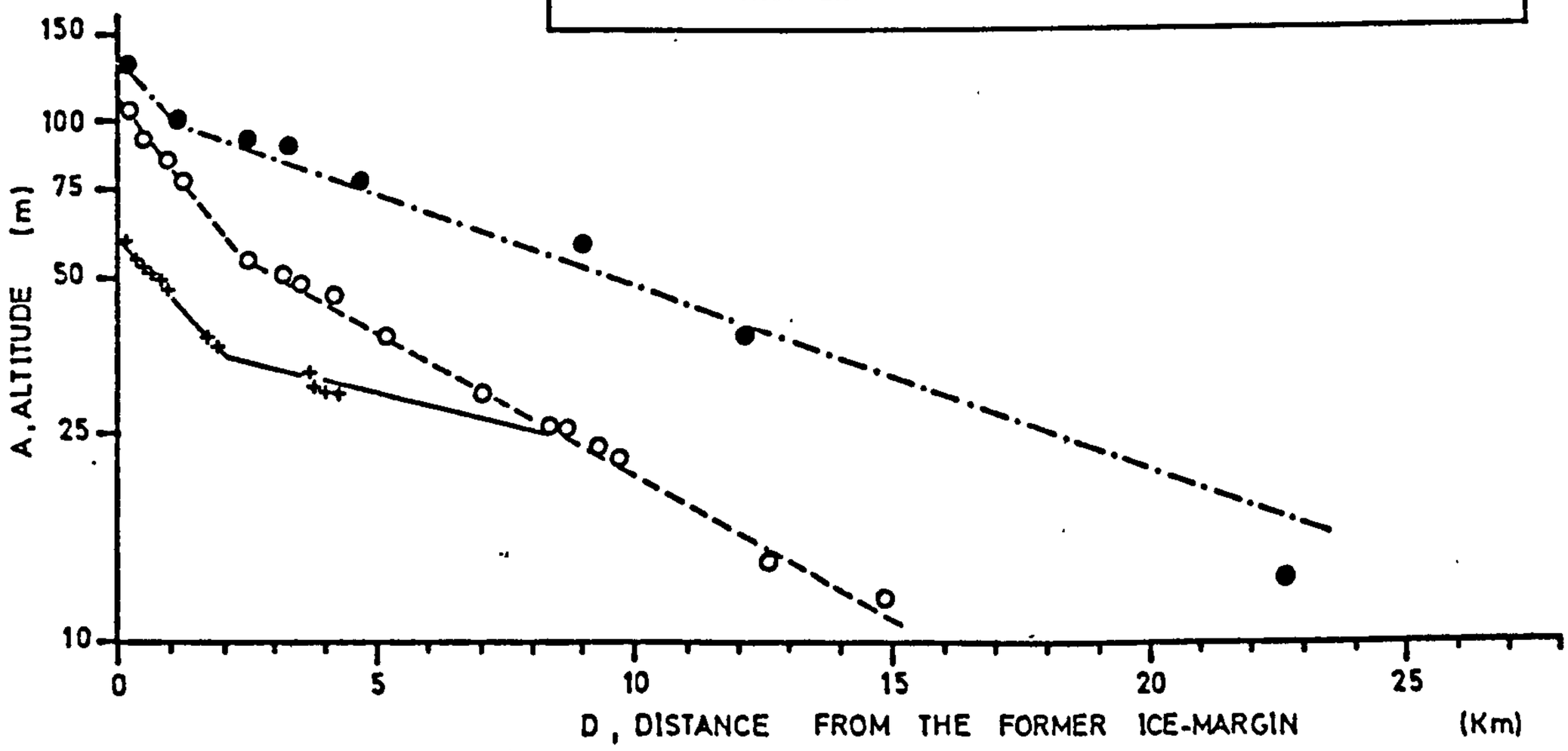


Figure 4.19
Former longitudinal profiles of the Scottish outwash deposits

		PROXIMAL SLOPE	DISTAL SLOPE
—·●—·	South Esk	$A = 130 \cdot 2\, D^{-0 \cdot 11}$	$A = 120 \cdot 2\, D^{-0 \cdot 04}$
—-○-—	North Esk	$A = 108 \cdot 9\, D^{-0 \cdot 12}$	$A = 74 \cdot 9\, D^{-0 \cdot 05}$
—+—	Westwater	$A = 61 \cdot 3\, D^{-0 \cdot 12}$	$A = 38 \cdot 7\, D^{-0 \cdot 03}$



similar degrees, namely between 0.34 and 0.40 m/km. The percentage increases in gradient relative to the original gradient, however, have been quite considerable over the more gently-sloping distal reaches (see the last two columns of Table 4.3).

The extrapolation of shoreline isobases upstream across paleo-outwash surfaces is a method which can reliably be employed only where the outwash deposits can be traced directly into the shoreline terrace (see Stephens and Synge, 1966). Nevertheless, the levelling of shorelines, plotting and extrapolation of isobases, and interpretation of the relation of the isobases to a true and extensive outwash surface all contribute to variability in the calculation of gradient. These sources of error, in this study, have tended to act towards minimising the differences in gradient between the former and present profiles of the paleo-outwash surfaces.

4.7. POST-DEPOSITIONAL MODIFICATIONS: II. ALTERATIONS TO THE INTERNAL STRUCTURE AND COMPOSITION OF THE OUTWASH DEPOSITS

4.7.1. INTRODUCTION

Changes in gradient over a prolonged period of time can affect large portions of the outwash body as a whole, but there is also evidence of localised, post-depositional alterations which have little morphological expression on the land surface. The post-depositional processes include those that have operated in association with cold climatic conditions (e.g. in a periglacial environment), in which ground ice activity is structurally important, and those that have operated throughout post-glacial time, such as the development of a soil cover, and the progressive leaching and subsequent accumulation of minerals by the action of percolating groundwaters. The older the deposit, the more likely it is to exhibit alterations in structure and composition which have resulted from post-depositional processes. It is clearly important to identify and to

account for these kinds of alterations when making a sedimentological comparison of contemporary and paleo-outwash deposits. The absence of particular features in a contemporary deposit, for example, may not reflect different depositional or initial chemical conditions, but subsequent structural deformations or progressive chemical weathering in the paleo-deposit.

4.7.2. INTERNAL STRUCTURAL DEFORMATION OF SEDIMENTS

The original sedimentary bedding of outwash deposits may be locally disrupted by the development of ice-wedges and involutions. These periglacial structures have been identified in many areas of eastern Scotland, chiefly by Fitzpatrick (1958) and Galloway (Common and Galloway, 1958; Galloway, 1961a,b; see also Rice, 1959a) and more recently, by other workers (e.g. McManus, 1966; McLellan, 1969b). Fossil ice-wedges, and amorphous and crude festoon-like involutions (West, 1968) have been recorded from many sections exposed in the North and South Esk outwash deposits. The present author observed seven wedge 'pseudomorphs' within an exposed distance of only 250m in the North Esk gravel pit (e.g. figure 4.20, page 211 and see figure 5.26, page 294), while 'frost-festoons' and other periglacial contortions were visible to a depth of about 0.60 - 1.00m even in the coarse outwash deposits (figure 4.21). The effects of periglacial ground ice activity in modifying the original position and orientation of coarse sedimentary particles are discussed in Chapter 6.

The structure of the deposit may also be disturbed, at least near the ground surface, by the development of soil and the associated growth and activities of plant roots and soil micro- and macro-organisms (figure 4.22).

4.7.3. MODIFICATION OF SEDIMENTARY PARTICLES

The major chemical changes affecting the particles of a sedimentary deposit result from prolonged groundwater activity. The downward percolation of groundwater leaches

FIGURE 4.20
ICE-WEDGE PSEUDOMORPH IN THE NORTH ESK GRAVEL PIT (E13)

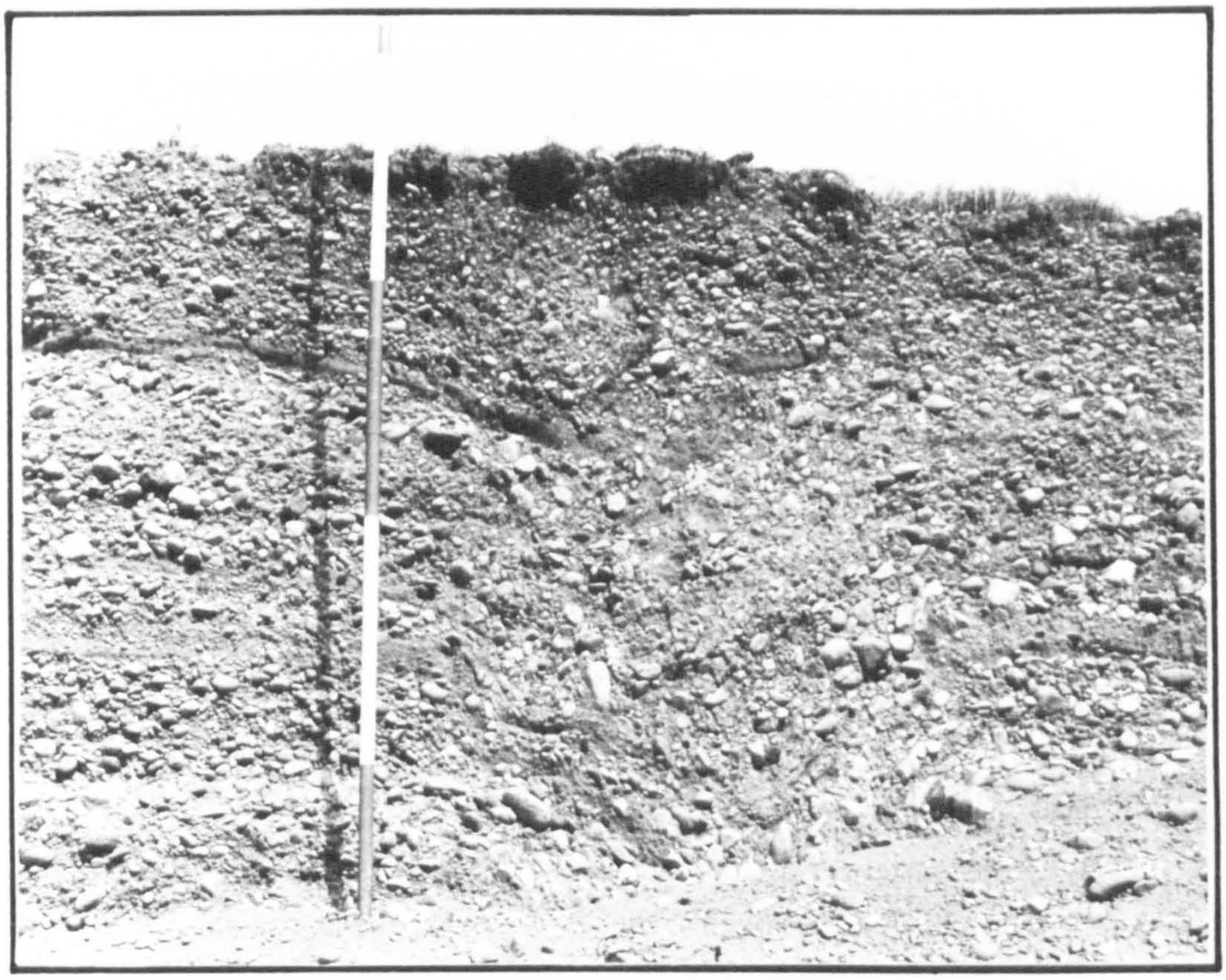


FIGURE 4.21
PERIGLACIAL CONTORTIONS IN THE NORTH ESK OUTWASH DEPOSITS

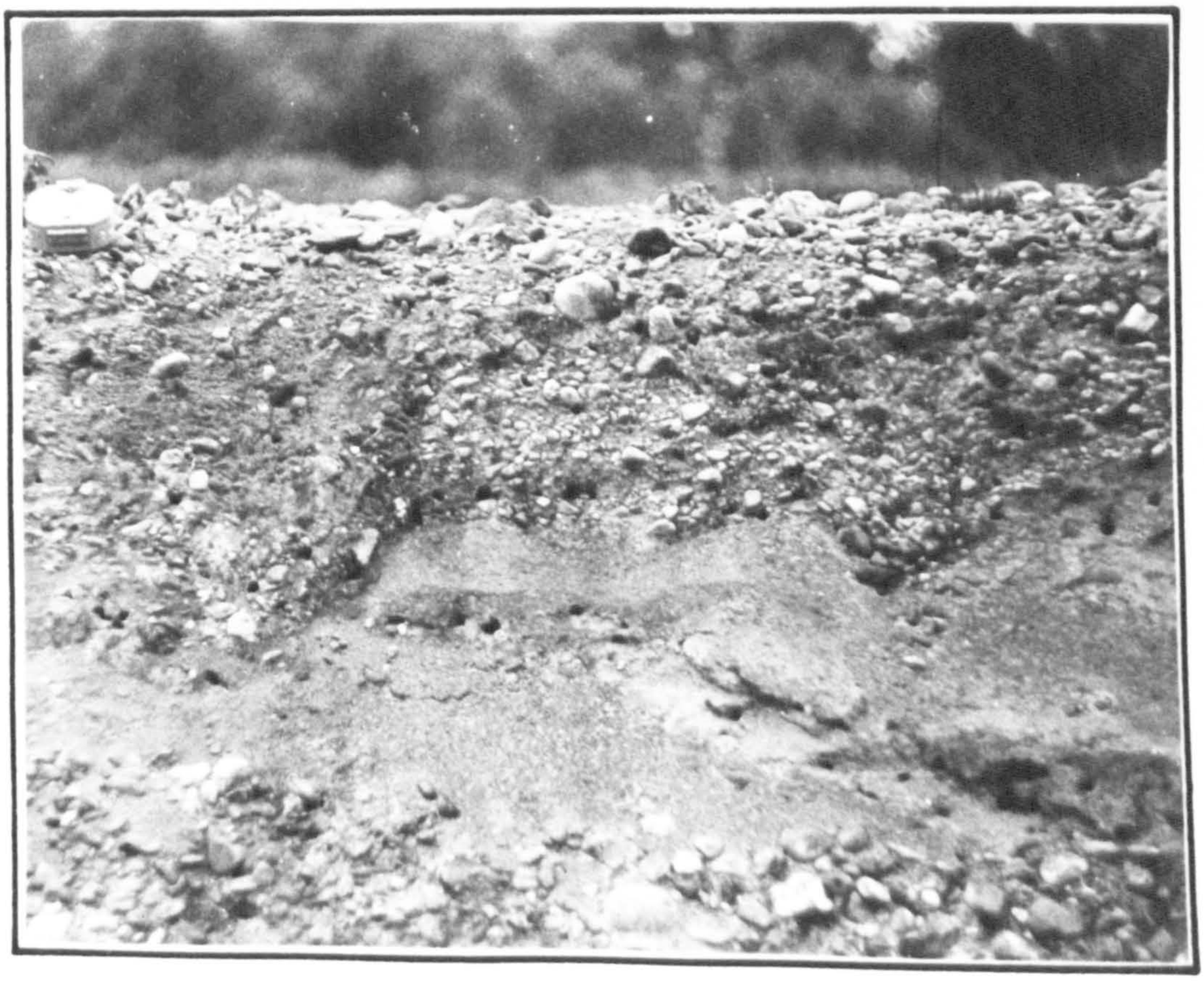


FIGURE 4.22 EXAMPLES OF SOIL DEVELOPMENT AND CHEMICAL ALTERATION OF COARSE, NORTH ESK OUTWASH DEPOSITS

the upper horizons of its soluble elements, in particular iron and manganese. The ferro-manganese solutes are then redeposited in the lower, 'B-horizon', at a level where groundwater ceases to penetrate to greater depths. The concentrated accumulation of iron in particular, results in the development of a 'hard-pan'. Fitzpatrick (1956, 1958) considers that this compact layer, or 'indurated horizon' represents the fossilised permafrost layer, lying 40-60cm below the ground surface. The depth of the 'active layer' corresponds with the depth to which involutions have been observed (see previous section). A zone of particles which are highly discoloured by the formation of coatings of oxides of iron (orange discolouration) and manganese (black discolouration) (figure 4.22) may mark the fossil 'permafrost-table' (Fitzpatrick, 1956). Romans et al. (1966,1973) have examined soils both from the upper reaches of Glen Esk and from the lowland outwash deposits near the North Esk gravel pit. They consider (1973) that the manganese coatings indicate the modifying action of a rising watertable during the last wet climatic phase of the Sub-Atlantic period.

Measurements of sediment reaction indicated that the groundwater is relatively acid. The average of over fifty pH recordings of the North Esk outwash deposits was less than 4.0. Acid groundwater is an effective solvent agent, and particles can become pitted, corroded, and rotten (see Chapter 6.5).

Oxidation and hydration are also important weathering processes, whose effects become more dramatic with the increasing age of the deposit. Coatings of iron oxide and the development of 'weathering rinds' are two features resulting from oxidation, while hydration results in the granular disintegration of polymineralic clasts. Measurements and interpretation of some of the particulate effects of post-depositional alterations in the contemporary and the paleo-deposits are considered in greater detail in Chapter 6.

4.8. COMPARISON OF METHODS OF SEDIMENT SAMPLING IN THE BOSSONS AND SCOTTISH DEPOSITS

4.8.1. SELECTION OF SAMPLING SITES

Methods of selecting sites for sediment sampling necessarily vary between areas whose whole surface is exposed and areas which are exposed only in scattered localities. Some of the problems involved in (a) choosing sites and (b) comparing sediment data based on two different sampling systems are discussed by Krumbein (1959a), who states that if the sampled population can be considered to be statistically and geologically equivalent to the target population, objective generalisations, inferences and predictions can be made. Krumbein also states that if there is no free choice of sampling site (the 'upper level of sampling'), it is important that selection of the samples themselves (the 'lower level of sampling') is based as far as possible on probability sampling methods. Random probability sampling was carried out for both the selection of sites and for the samples at the site in the proglacial valley of the Bossons glacier, but, because of lack of choice of sampling site, probability sampling was carried out for only the samples themselves in the paleo-deposits.

4.8.1.1. Bossons proglacial valley-train

The random selection of sites on the valley-train was subject to only one limitation, for it was decided that the 'conceptual population' of sediments to be sampled were those which represented, as far as possible, deposition within a range of high flow regimes. Sample sites were thus restricted to the exposed surface of a shoal or bar (Plate 4), and any site selected which would have been situated on a boulder, in an area of vegetation, in a quicksand or kettle-hole, or in a meltwater stream, was rejected. Sediment samples would not have been directly comparable with those from the paleodeposits if they had

been obtained from all possible sites of all kinds, because only the coarsest sediments were sampled from the ancient outwash deposits (see below). Samples selected from stream beds would not have been directly comparable with paleo-deposits either, because the sediments would represent only some of the particles actually in motion, rather than the 'final' sedimentary deposit (Church, 1970).

A major disadvantage, however, is that it may be impossible to obtain subsurface sediment samples from an active outwash deposit. The water-table is very high (see Chapter 3) during the meltseason, the ground is frozen hard in winter, and few natural sections are present if the meltwater streams are aggrading. Subsurface samples were obtained only to (a) a depth of 1.20m from the walls of kettle-holes before infilling with meltwater and (b) a depth of 10cm in a series of pits excavated in April 1971.

Each random sampling plan produced several pairs of sites situated adjacent to each other on the shoal surface; the samples from these sites were used to determine the degree of local sediment variability.

4.8.1.2. The four Scottish paleo-deposits

Samples were selected from all the available outcrops in the outwash gravels (see figure 4.23, page 216) (Pettijohn, 1962), and the number of samples was therefore limited by the number of exposures. Augering or the construction of excavations at other sites was not considered feasible because of the coarse nature of the deposits and the thickness of the overlying soil and disturbed 'gravel' horizons. The exposures of outwash deposits were therefore limited to commercial gravel pits, and road and river cuttings. The resultant distribution of sites is very irregular, particularly in the relatively poorly exposed South Esk deposits. The number of sites is, however, sufficiently large, since the addition of new sites does not significantly alter the pattern of many of the measurements already obtained (Pettijohn, 1962). It may, therefore, be assumed that

FIGURE 4.23
TWO TYPICAL EXPOSURES IN THE NORTH ESK OUTWASH DEPOSITS

(a) Site E24

(b) Site E33

the sampled population approximates the true 'geological population' (Krumbein, 1959a). Sites were identified from (a) large-scale Ordnance Survey maps, and (b) field coverage of the outwash areas. Over 130 sections were examined in the four outwash series (Plates 5 and 6).

4.8.2. SELECTION OF SAMPLES AT SITE

Two types of sediment samples were collected:
(1) Aggregate samples, or sediment samples which comprised the whole range of particle sizes present;
(2) Coarse particle samples, or samples comprising at least 50 particles, each exceeding 16mm in diameter (-4 phi on the Wentworth grain-size scale; see Krumbein, 1938; Krumbein and Sloss, 1963; Griffiths, 1967).

4.8.2.1. Bossons proglacial valley-train

(1) Aggregate samples of surface deposits were scooped up from a 25cm square marked out on the shoal surface at each site. Samples weighed about 0.5kg. Subsurface samples, selected from 10cm below the surface and just above the water-table at the time of sampling, were also collected in order to determine any local vertical variability in sediment characteristics, and to estimate whether the subsurface sediments had been modified since deposition, such as through the removal of the fine fractions by ground water flow. The area covered by the sample sites represented about 0.2 per cent of the total shoal area of the valley-train at the time of sampling. Aggregate samples were also collected systematically from the lateral and terminal moraines (see also Chapter 2.6).

Mechanical analysis was carried out on over 230 aggregate sediment samples using sieves at $\frac{1}{2}$-phi intervals (Krumbein, 1938; Krumbein and Pettijohn, 1938; Krumbein and Sloss, 1963; Griffiths, 1967). Sieve analysis can give rise to errors in measurement and in interpretation if inaccurate mesh sizes are used and if clogging with fine

fractions occurs, thus increasing the variability of results. Ten per cent of samples were re-sieved to provide an estimate of sieving errors. The intermediate diameter of particles governs the sieve size distribution which is arbitrarily divided into a series of size classes, but analysis based solely on the intermediate diameter does not necessarily reflect the behaviour of the particle when it was being deposited in a fluid. Measurement of the 'settling velocity' of the grains would have provided a more accurate sediment 'size' distribution, for it would also reflect particle shape and specific gravity (Baker, 1920). However, sieving is a quicker method, providing results which are directly comparable with data obtained by most other workers. Hydrometer analysis was only carried out to measure the proportion of silts and clays (< 0.063 mm, > 4 phi) present. This method was not considered sufficiently accurate to measure the small quantities (< 5 per cent) of fine fractions in the outwash sediments.

(2) Over 100 coarse particle samples have been collected from the valley-train over the period of study. The sediment at each sample site was scooped up and quartered. One pile was requartered and all coarse particles extracted for measurement. If there were less than 50 pebbles in the first quarter, the coarse particles from the second pile were extracted, and so on. A small-scale study of a single shoal entailed the collection of over 40 samples each consisting of the ten largest clasts.

Selection of the coarsest particles from tops of bars is based on the premises that (a) the sediment samples reflect the maximum flow velocities transporting the sediment to the site of deposition (see Bradley et al., 1972); and (b) the distribution of coarse particle sizes is statistically related to the size distribution of the whole sediment (Krumbein and Lieblein, 1956; Pelletier, 1958). Many other workers have employed the method of measuring the coarsest particles in an alluvial sediment

(e.g. Kreuter, 1913; Jones and Quam, 1944; Inderbitzen, 1959; Bluck, 1964; Ehrlich and Davies, 1968; McDonald and Banerjee, 1971; Bradley et al., 1972; Stanley and Wayne, 1972; Boothroyd, 1972). The measurement of the coarsest clasts at each site can then prove valuable, firstly for relating the sediment sizes to the highest meltwater discharges that deposited material at the site; and secondly for relating the sediments at one site to those at upstream and downstream sites at which maximum sediment sizes, deposited by the highest meltwater discharges, have also been sampled. The largest particles at each site are more likely to be 'equivalent' and directly related, than if samples are derived from sediments in a wide variety of depositional 'facies'. Local variability in maximum particle size may result from variations in local elevation (Church, 1972), subsequent removal of fine fractions by water or wind, addition of material by water, wind or from valley-side slopes (Chapter 3), or from the lack of sufficient sampling sites. The results of these local effects can generally be determined from size analyses, and they do not reduce the significance of downstream trends in size characteristics (see Chapter 7).

4.8.2.2. <u>The four Scottish paleo-deposits</u>

(1) Discrete, or 'spot' <u>aggregate samples</u> (Krumbein, 1938) were extracted from each 'sedimentation unit' (Otto, 1938) in the exposed section of outwash deposits. Otto defines the sedimentation unit as an apparently homogeneous layer of a sediment which was deposited under 'essentially constant physical conditions'. More recently, work by Jopling (1963,1964) indicates that 'essentially constant conditions' can include a fairly wide range of hydrodynamic conditions, which results in the deposition of distinctive laminations. A contiguous vertical series of laminations occurring in fine deposits, therefore, frequently formsa single sedimentation unit. Each distinctive sedimentary stratum represents deposition under a particular, and small range of, stream flow and bed material conditions.

Vertical variations in the size composition of a sedimentation unit thus reflect changes of stream facies through time (see below, and figure 4.24).

(2) Coarse particle samples were selected from the coarsest sedimentation unit, or units, at each vertical section. All particles were extracted from a relatively homogeneous stratum, and placed on the ground in a pile. The same particle sampling procedure was carried out as at Bossons (see above). A small-scale study of local vertical and downstream particle size variability was undertaken on a 300m x 2m gravel pit face (North Esk gravel pit). The ten largest clasts were selected at vertical intervals of 20cm, and at 20m intervals along the pit face (apart from the initial 8 sections which lay 10m from each other)[1]. Samples of the ten largest particles were also selected from the outwash deposits in Glen Lethnot, for they were too large (up to 1m diameter) to be removed for measurement elsewhere.

The sediments forming paleo-outwash deposits are distinguished from the actively forming deposits of a present-day proglacial environment in that they are 'time-transgressive' (McDonald and Banerjee, 1971). The coarsest deposits represent the preservation and progressive burial of what was probably a former bar surface (Allen, 1970b; Rust, 1972) or channel bed. A vertical section thus records variations in sedimentological conditions over a period of time. A series of adjacent vertical sections therefore represent sediments deposited at a variety of locations at different moments through time, whereas the present-day outwash surface represents sediment variability through space but generally deposited within a single 'period' of time. The whole thickness of deposits may include evidence of a wider or narrower range of facies than is present on a contemporary outwash surface, according to the degree of preservation and the local facies change through time in a single locality. A vertical series of outwash sediments is also likely to contain localised breaks in the depositional sequence, as channels were temporarily abandoned or as stream erosion occurred, but both potentially

1 See Plate 7

Figure 4.24
Laminations in a single sedimentation unit
and a vertical change from fine to coarse sediment facies

depositional and non-depositional facies can be observed in present-day deposits (see discussions in later chapters).

The samples of coarse particles at successive sites, therefore, represent downstream variability in maximum discharge conditions depositing debris at the sample site at any time during the formation of the vertical sequence under investigation. The debris was subsequently buried by finer deposits from overbank flow or lesser flood or sheet flows. The coarse sediments from both the contemporary and the paleo-outwash deposits have thus originated from maximum flow conditions, and are considered to be sufficiently 'equivalent' to allow direct and valid comparison of particle size characteristics.

4.9. CONCLUSION

The comparison of the physical backgrounds to proglacial outwash deposition has provided valuable information relating to the nature of outwash formation and of sediment characteristics. The comparison has also proved to be worthwhile in that each region investigated has demanded a widely differing approach to observation and data collection, especially in terms of differences of scale, morphology and sampling. This part of the study has led to greater understanding of the wide variety of methods, problems and possible interpretations associated with the morphological, stratigraphical and sedimentological analysis of several series of outwash deposits.

Chapter 5

THE FORMATION OF OUTWASH DEPOSITS: MORPHOLOGY, PROCESSES AND SEDIMENTARY STRUCTURES OF A BRAIDED FLUVIOGLACIAL ENVIRONMENT

Additional list of symbols

A	Volume of sediment removal (Corbel's formula, 1959)
a	Reference depth in Rouse's sediment diffusion equation
B.I.	Braiding index (Brice, 1964)
B_L	Bar length
B_W	Bar width
C	Chezy's friction coefficient; sediment concentration; constant in settling velocity equations
C_W	Channel width
D	Mean particle diameter; Rock density (Smith, 1972)
d	Channel depth
E	Depth of stream, or Total precipitation - total evaporation (Corbel, 1959)
F	Width-depth ratio (Schumm, 1960a)
F_r	Froude number
g	Acceleration due to gravity
k	Von Karman's constant
L	Number of channel links; Length of reach (Brice, 1964)
L_b	Length of bar (Brice, 1964)
L_c	Total length of channels
L_v	Length of valley axis
M	Index of percentage of silt and clay (Schumm, 1960a)
N	Number of channels; Number of nodes
$\bar{N}$	Mean number of channels
n	Manning's friction coefficient
P	Number of bifurcation points
Q_w	Total annual meltwater discharge
q_s	Mean discharge of sediments in solution
R	Nearest neighbour statistic
R_h	Hydraulic radius (also given as R_s)
$\bar{r}_a$	Mean distance between nodes
S	Channel sinuosity
$\bar{S}$	Mean channel sinuosity
S_B	Percentage of silt and clay in channel bank (Schumm, 1960a)

S_c	Percentage of silt and clay in channel (Schumm, 1960a)
S_d	Total amount of sediment deposited on valley-train
S_g	Total amount of sediment removed from glacier basin
S_w	Total amount of sediment removed from valley-train
T	Total discharge of sediments (Corbel, 1959)
T_c	Total amount of $CaCO_3$ in solution (Smith, 1972)
T_m	Total amount of $MgCO_3$ in solution (Smith, 1972)
V_*	Shear velocity (also given as U_*)
V_s	Settling velocity of particles (also given as v)
V_u	Upward velocity component
w	Channel width
X	Amount of sediment removed in solution (Smith,1972)
y	Selected depth in Rouse's sediment diffusion equation
γ	Specific weight of fluid
η	Viscosity of fluid
ρ	Density of fluid
τ_c	Tractive force

Chapter 5

THE FORMATION OF OUTWASH DEPOSITS: MORPHOLOGY, PROCESSES AND SEDIMENTARY STRUCTURES OF A BRAIDED FLUVIO-GLACIAL ENVIRONMENT

5.1. INTRODUCTION

The identification and morphological analysis of a series of proglacial outwash deposits are only the first stages in the interpretation of an environment which may comprise a wide range of sediment types. Proglacial outwash sediments have generally been deposited by streams flowing in a braided channel system (see Chapter 1.1.). Investigation of the morphological features of this characteristic stream pattern, and of the channels which make up the pattern, is also important. The channel pattern and form both result from a sequence of hydraulic mechanisms, and once the channels are initiated, give rise to particular fluvial processes operating within them. Evidence of channel pattern, channel morphology and the processes of erosion, transport and deposition of sediments observed in present-day meltwater streams can often also be recognized in paleo-outwash deposits. Comparison with present-day empirical data may allow many paleo-flow parameters to be estimated both qualitatively and quantitatively.

5.2. POSSIBLE CAUSES OF A BRAIDED OUTWASH SYSTEM

A braided stream network is composed of multiple interconnecting channels, only separated from each other at low and moderate flows by islands or mid-channel bars of alluvial sediment. Many workers have tried to account for the initiation of a braided, or 'anastomosing' channel system, in which streams continually divide and rejoin.

5.2.1. EXCESS SEDIMENT LOAD

Most support in the literature for the origin of braiding is given to the apparent overloading of the stream with sediment (e.g. Hjulström, 1952; Smith, 1970; Church, 1972). 'Overloading' means that part of the sediment load entering the stream reach is too great to be removed by the

water flowing through the reach at that time. The stream attempts to maintain hydrodynamic equilibrium by selective deposition of its excess load. The coarsest particles are dropped first to form a 'mid-channel bar', which continues to be built up towards the water surface. Eventually the bar directs the stream flow into two single channels, each of which is of a more efficient shape and size to remove the sediment load than the former, single channel (Mackin, 1948; Krigström, 1962; Church, 1972) (see section 5.3.). The possible causes of stream 'over-loading' are more relevant to proglacial environments than are the possible causes of braiding, because of the high rate of sediment production from many glaciated basins.

5.2.2. CHANNEL SLOPE

Channel slope, or the hydraulic gradient, has also been suggested as a possible cause of braiding (Smith, 1970). Several workers have found that a braided stream network occurs on a steeper slope than a meandering stream with a similar discharge of water (Leopold and Wolman, 1957; Leopold, Wolman and Miller, 1964; Wolman, 1967; Schumm, 1971). The steeper gradient associated with a braided system results in greater flow velocities, and therefore greater ability to transport coarser debris. Coarser stream loads are more likely to be deposited as excess load with periodic falls in stream discharge, and to be resistant to erosion once they have been deposited on the channel floor. However, a steeper gradient is not necessarily associated with a braided reach. McDonald and Banerjee (1971), for example, found that braiding occurred on the gentler slopes of the distal reaches of the valley-train while the steeper proximal slope of the valley-train was being actively incised. Local stream incision was also observed on the proximal slope of the Bossons valley-train.

The actual channel slope may not necessarily produce sufficient overloading to result in stream bifurcation. A change in slope, therefore, may be of greater significance. A temporal or downstream change in gradient causes changes

to occur in other channel parameters, such as width and depth, and in hydraulic parameters, such as bed shear stress, friction, turbulence and velocity (see Mackin, 1948). These changes could easily result in a new sedimentary environment, such as that of a braided stream system.

5.2.3. COARSE, NON-COHESIVE BANK MATERIAL

The presence of coarse debris is an additional and important factor in the development of 'overloading' (e.g. Leopold, Wolman and Miller, 1964). Deposits of coarse debris not only provide resistance to stream flow (5.2.2.) but they also frequently form poorly-sorted, non-cohesive banks. The saturation of these bank deposits during rising flow stages causes undercutting, slumping and 'sloughing' (Lane, 1937) and caving of bank material. The rapid erosion of banks can provide abundant, regular inputs of debris to the stream. The additions of debris may either be transported farther downstream before being deposited, or may be selectively eroded at the point of entry, eventually forming a 'lag' deposit or core of a bar which accumulates around it (Leopold, Wolman and Miller, 1964; Dury, 1969).

5.2.4. VARIATIONS IN DISCHARGE

Braided channels commonly occur in areas characterised by ephemeral stream flow, such as proglacial and semi-arid regions. Mean annual discharge is usually low, but the annual régime is marked by sporadic, high discharge or flood events. Several workers have, therefore, regarded large and rapid variations in stream discharge as an additional cause of braiding (e.g. Doeglas, 1962; Leopold, Wolman and Miller, 1964; Blench, 1969; Selley, 1970; Smith, 1970; Nordseth, 1973a). Fluctuations of stream discharge and of water level can have several important results. Firstly, the rate of bank erosion is accelerated through repeated flooding and saturation so that the stream becomes more heavily loaded, and the channel migrates into adjacent alluvial deposits (e.g. Klimek, 1973). Secondly, the redistribution of sediments from the whole deposit is also

accelerated during high discharge conditions (Smith, 1970). The stream deposits become relatively unstable because of rapid changes in particle location, even though the distance of each movement may be quite small (Church, 1972). Thirdly, frequent falls of flow stage of a heavily-laden stream mean that local sediment aggradation is also a frequent event. Mid-channel and/or overbank deposition may occur during the decline of stream competence following each peak flow, and the likelihood of stream bifurcation is consequently increased.

Fluctuations in discharge promote sediment mobility and channel instability, but only where the sediments are initially noncohesive, and possess a low threshold of erosion. Large and rapid variations of discharge are, therefore, not essential for the initiation of a braided channel system.

5.2.5. SELECTIVE SORTING

A stream which does not possess an excess load may also exhibit a braided channel pattern. Stream branching may result from the selective removal of sediment from the channel floor, leaving behind a coarse 'lag' deposit, around which the channel is deepened by scour (section 5.3.1.). The 'lag' deposit eventually rises above water-level as scouring continues or as the water-level falls (Leopold and Wolman, 1957; Brice, 1964).

5.2.6. OTHER CONTRIBUTORY FACTORS

The absence, or near absence of a vegetative cover may also enhance the likelihood of a braided channel network developing (Doeglas, 1962; Brice, 1964) by not providing a stabilizing root system on exposed bar or bank deposits (Tricart, 1970), and by allowing high rates of surface run-off (Selley, 1970). Minimal delay in surface run-off from the 'interfluves' or banks subsequently results in greater fluctuations in stream discharge. Surface run-off would be even more rapid if the subsoil were impermeable (Doeglas, 1962).

5.2.7. CONCLUSION

The 'overloading' of a stream is one of the primary causes of braiding, but many factors can act to produce an excessive sediment load. The braided channel system of the Bossons valley-train, for example, appears to have resulted from stream overloading in a generally aggrading system.

The excess sediment load has resulted from a sudden downstream decrease in gradient (see Chapter 3), a series of coarse, unstable bank deposits (see Chapter 6), rapid and large variations in meltwater discharge (see Chapter 2.5.), and a relative absence of a vegetative cover (see Chapter 2.6.). Similar conditions are also likely to have been present during the formation of the Scottish paleo-outwash deposits.

5.3. CHANNEL PATTERN IN A BRAIDED SYSTEM

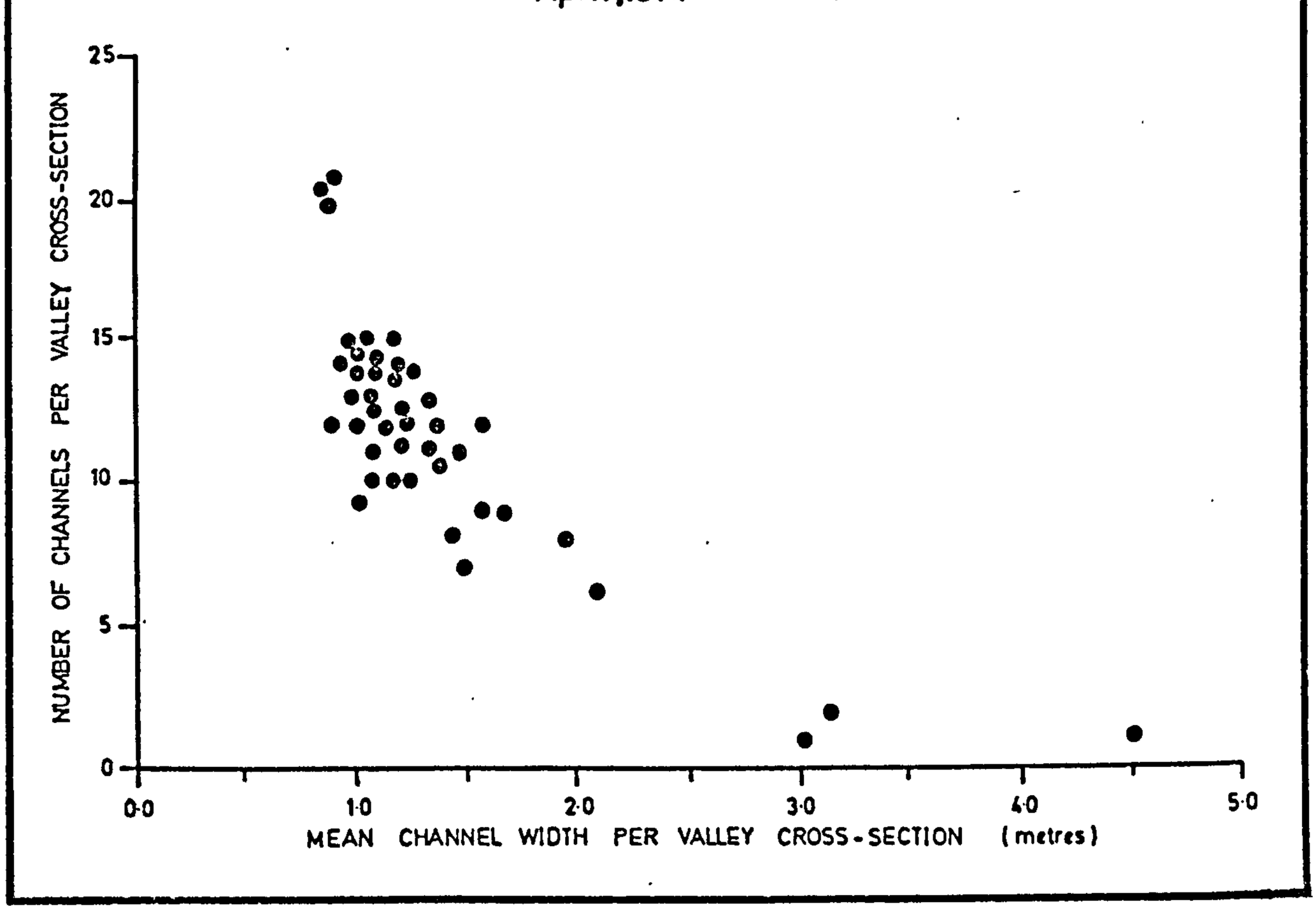
Description and measurement of a braided stream network has been relatively neglected in the literature, partly because channel migration can be so rapid, and partly because the large number of stream channels, which are frequently present, are difficult to measure. It is important, however, to derive some measure or index of braiding, in order to establish the degree to which a braided stream network has developed and to determine how such a pattern can change through time.

5.3.1. MEASUREMENTS OF A BRAIDED CHANNEL SYSTEM

5.3.1.1. The number of channel segments is a useful measure of the degree of multiplicity of channel development. The number of channel segments is related to several indicators of the hydraulic environment: (a) flow stage; (b) the amount of mid-channel aggradation which is itself affected by sediment load, slope and discharge (see section 5.5.); and (c) the presence of valley constrictions (see also Fahnestock, 1963). Fewer channels will be present where there are high flow stage conditions, little mid-

channel deposition acting to divert a single channel into two or more channels, and prominent valley constrictions. Higher flow stages result in fewer channels, for measurements on the Bossons valley-train show that the fewer channels there are, the wider are those channels (figure 5.1, page 231). Fewer, wider channels suggest that shallow, secondary channels have been flooded and that mid-channel aggradation has been insufficient to cause extensive channel bifurcation.

Figure 5.1
Mean channel width and the number of channels across the Bossons valley-train, April,1974



The number of channel segments on the Bossons valley-train (Table 5.1, page 233) rose by 45 per cent, (from 344 to over 500) between 1968 and 1973, despite a reduction of over 80m in valley length. The mean number of channel segments was considered to be a more comparative index of channel multiplicity because of the changing shape and area of the outwash body. However, the mean number of channel segments at each of 60 valley cross-sections (measured from the surveyed maps of the valley-train) also more than doubled between 1968 and 1973 (5.73 compared with 12.00) (figures 5.2 and 5.3). Overall valley-train aggradation has thus resulted in a measurable increase in the number of channel segments present and a corresponding decrease in the width of those channels (Figure 5.2) as a result of an increase in mid-channel deposition. The flow stage at the time of measurement has some effect on numbers of channel segments, but low flow stage is unlikely to be the sole cause of an increase in channel segment number because even in August 1971 the number of channel segments was large.

Valley constrictions have had an important effect on the number of channel segments on the Bossons valley-train. The constriction formed by the downstream valley outlet ensures that only a single channel ever forms in that locality. The former terminal moraine conspicuously reduced the number of channel segments lying across the valley-train in its vicinity in 1968 and 1969. Progressive burial of the moraine and extension of the outwash deposits across it since 1971 has caused the number of channel segments near the moraine to increase from 4 in 1968 to 14 in 1974 (Figure 5.3, page 235).

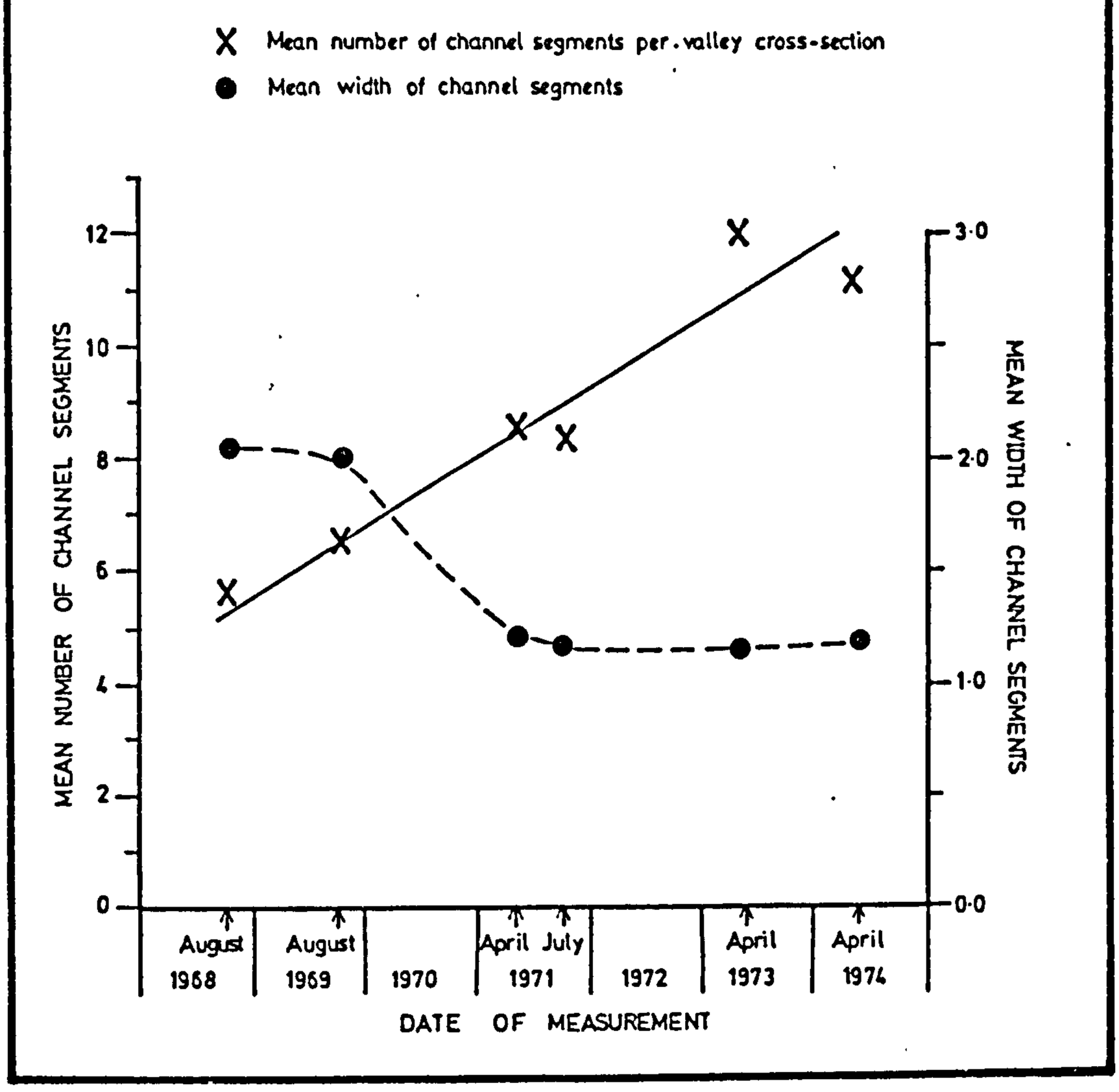
Table 5.1

Measures of channel pattern: Bossons valley-train 1968 to 1974

Measurement	Y E A R					
	August 1968	August 1969	April 1971	July 1971	April 1973	April 1974
Valley[1] length (m.)	297.60	300.00	229.00	222.00	216.00	212.10
Mean valley width (m.)[2]	33.64	33.56	36.26	37.63	41.79	42.68
Valley length/valley width	8.85	8.94	6.32	5.90	5.17	4.97
Valley area (m^2)	10011.26	10068.00	8310.80	8353.90	9026.60	9052.43
No. of x-sections[3]	60	60	49	47	44	44
No. of channel segments	344	396	431	406	528	500
Mean no. of channel segments[4]	5.73	6.60	8.80	8.64	12.00	11.11
Mean % channel area	34.57	39.26	29.71	26.99	33.54	32.49
Total channel length (m.)	1890.00	2358.00	2364.00	2316.00	2991.00	2796.00
Sinuosity[5]	6.35	7.86	9.99	10.43	13.85	13.07
Mean sinuosity[6]	1.11	1.19	1.14	1.21	1.15	1.18
No. of bars[7]	58	91	129	129	143	160
Mean bar length (m.)	19.06	14.47	11.20	12.28	13.99	11.87
Mean bar width (m.)	4.18	3.10	2.75	2.81	2.98	2.88
Mean bar length/width	4.56	4.67	4.07	4.37	4.69	4.12
Mean channel width (m.)	2.03	2.00	1.22	1.18	1.17	1.19
Mean bar orientation	338° 41'	341° 00'	332° 26'	332° 56'	325° 47'	327° 21'
Modal bar orientation	0°-9°	330°-339°	330°-339°	350°-359°	320°-329°	330°-339°
Valley-train orientation	342°	342°	330°	330°	328°	328°
Total no. of nodes, N	114	117	261	212	257	307
Mean no. of nodes, $\bar{N}$	1.90	2.95	4.35	4.51	6.32	6.98
No. of bifurcations, P	57	88	130	106	127	153
No. of links, L	171	267	393	318	382	462

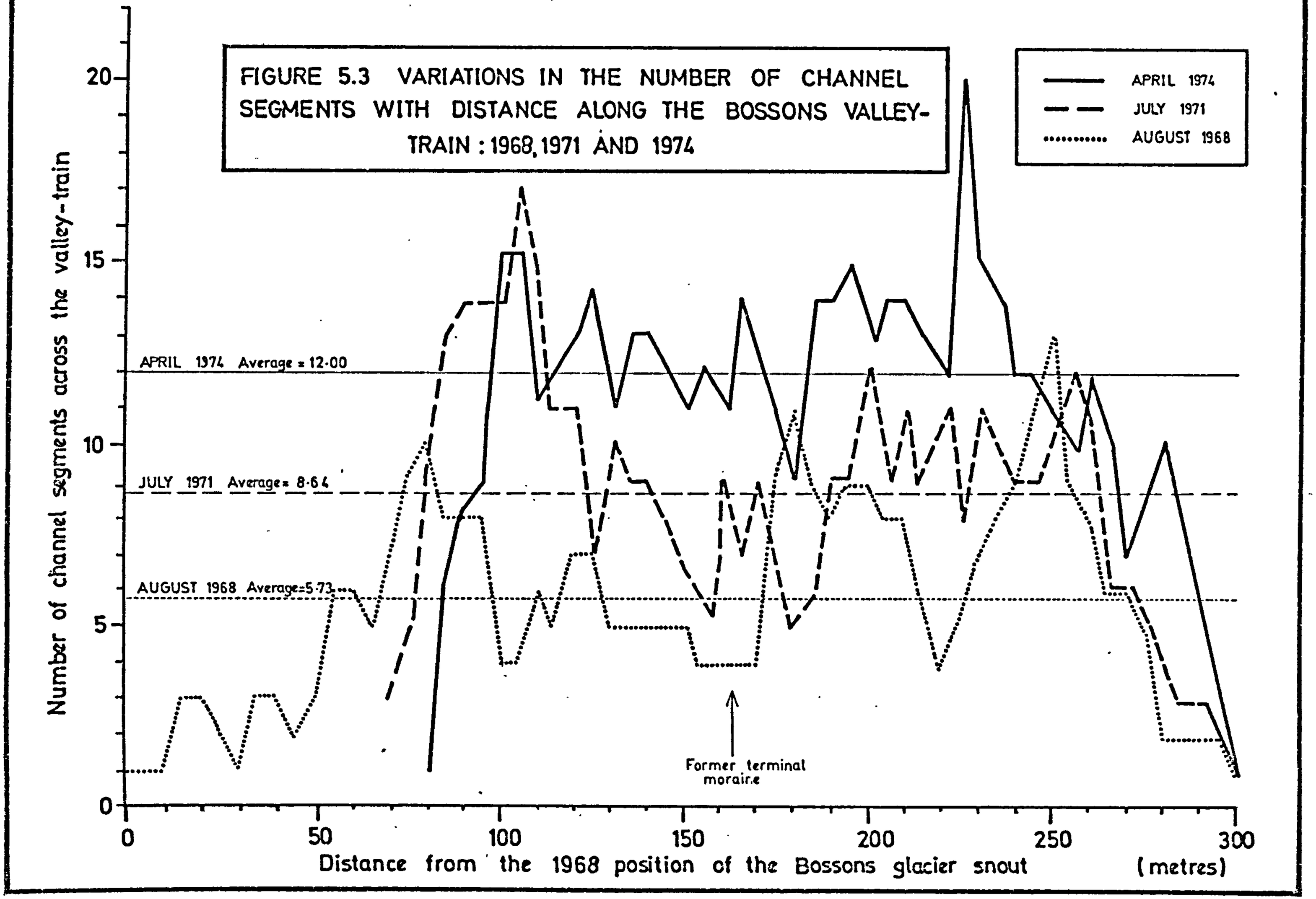
[1]'Valley' length is length of mid-valley axis along valley-train. [2]Mean valley width is average distance of each cross-section across valley-train to break of slope with lateral moraine. [3]Cross-sections were taken at 5m intervals along valley-train. [4]Mean number of channel segments per cross-section. [5]Channel length/valley length. [6]Sinuosity/mean number of channel segments per cross-section. [7] Midchannel bars only.

Figure 5.2
Changes in the mean number of channel segments and in mean channel width, Bossons valley-train, 1968 to 1974



5.3.1.2. 'Sinuosity', S, is defined as the ratio of total channel length to valley length (Leopold, Wolman and Miller, 1964). The greater the difference between the slope of the channel and the gradient of the valley, the greater the channel sinuosity (Schumm, 1968). Sinuosity values indicate, therefore, whether a stream channel is 'straight' ($S < 1.5$) or 'meandering' ($S > 1.5$). However, the index is generally applied to single channels, so that to take account of the multiplicity of channels, the mean sinuosity, $\bar{S}$, is probably more accurate:

FIGURE 5.3 VARIATIONS IN THE NUMBER OF CHANNEL SEGMENTS WITH DISTANCE ALONG THE BOSSONS VALLEY-TRAIN : 1968, 1971 AND 1974



$$S = L_c/L_v \qquad (5.1)$$

and $$\bar{S} = \bar{N}L_c/L_v \qquad (5.2)$$

where L_c is the total length of channels, L_v is the length of the valley axis, and $\bar{N}$ is the mean number of channels across the valley floor.

Sinuosity values are affected by the shape of the valley floor. The length-width ratio of the Bossons valley-train was about 9.0 in 1968 and 1969, with associated sinuosity values of less than 8.0 (Table 5.1, page 233). The length-width ratio fell as the glacier advanced, and by 1973 was about 5.0; the associated sinuosity index correspondingly rose to nearly 14.0 (Figure 5.4a and b, page 237). Sinuosity thus more than doubled between 1968 and 1974, indicating that total channel length has increased with increasing valley width (Figure 5.4c, page 237).

Mean sinuosity values, however, have not changed significantly through time (Figure 5.4a, page 237). Mean sinuosity has remained between 1.11 and 1.21, indicating fairly straight courses. These values are similar to those of 1.02-1.34 found by Williams and Rust (1969) on the braided outwash train of the Donjek River, Alaska. The relatively straight courses reflect the steeper gradients compared with those of meandering reaches (Leopold and Wolman, 1957), and the almost straight downvalley course followed by streams during flooding, and subsequently maintaining new, more direct channels.

5.3.1.3. Topological measurements

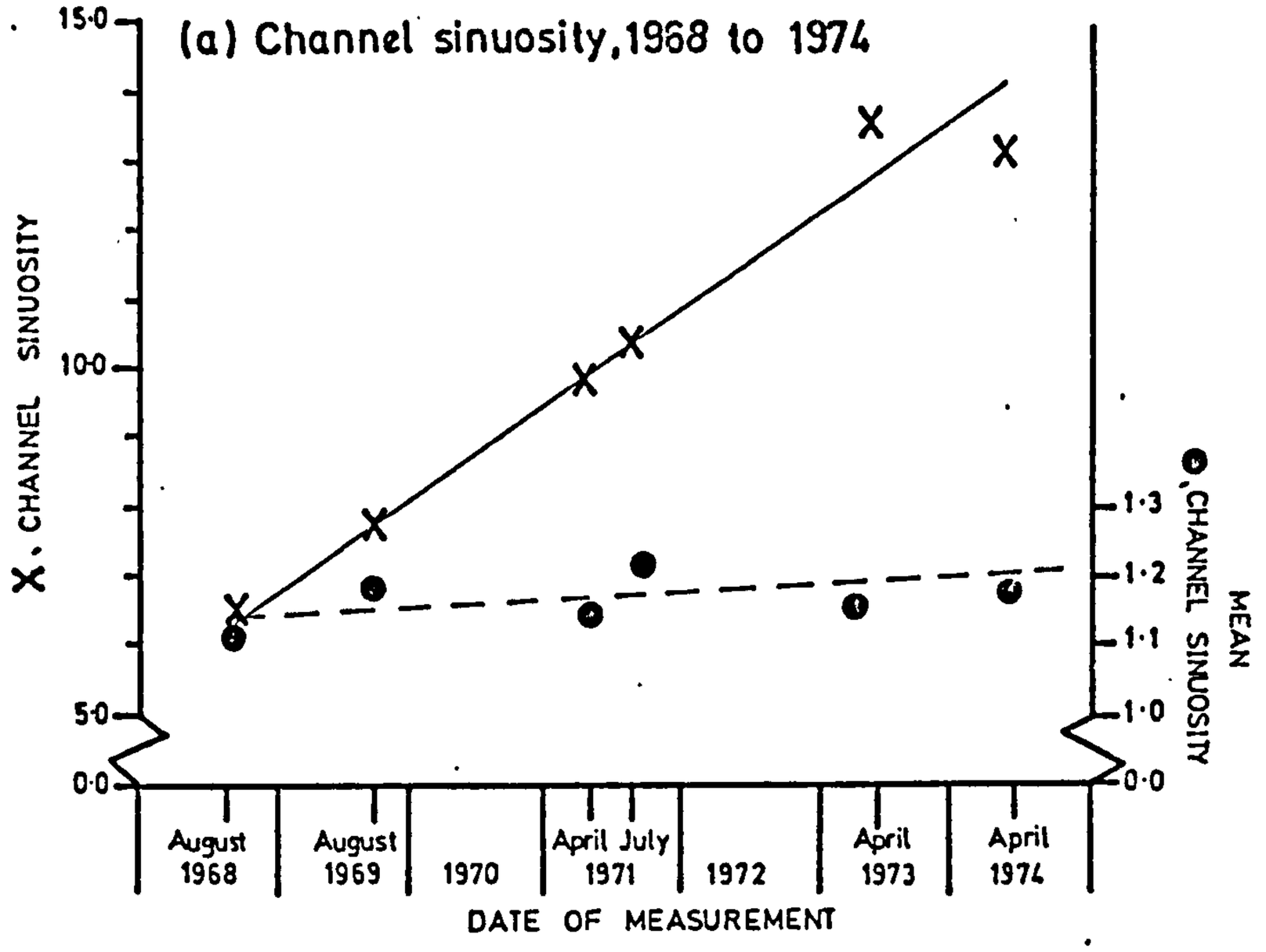
A more widely used and more rapid method of analyzing braided stream networks is based on the number of nodes, N (stream junction points), the number of links, L (stream segments between nodes), and the number of bifurcation points, P, using the equations:

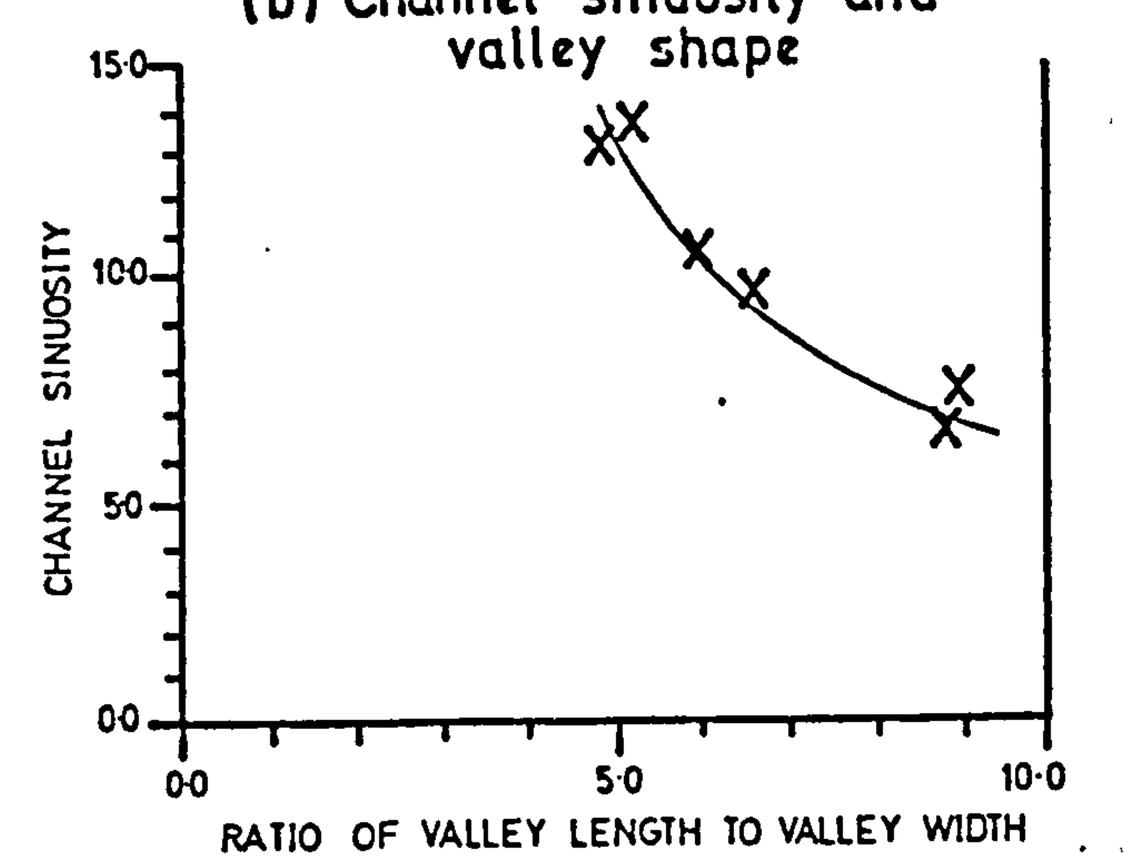
$$P = (N-1)/2 \qquad (5.3)$$

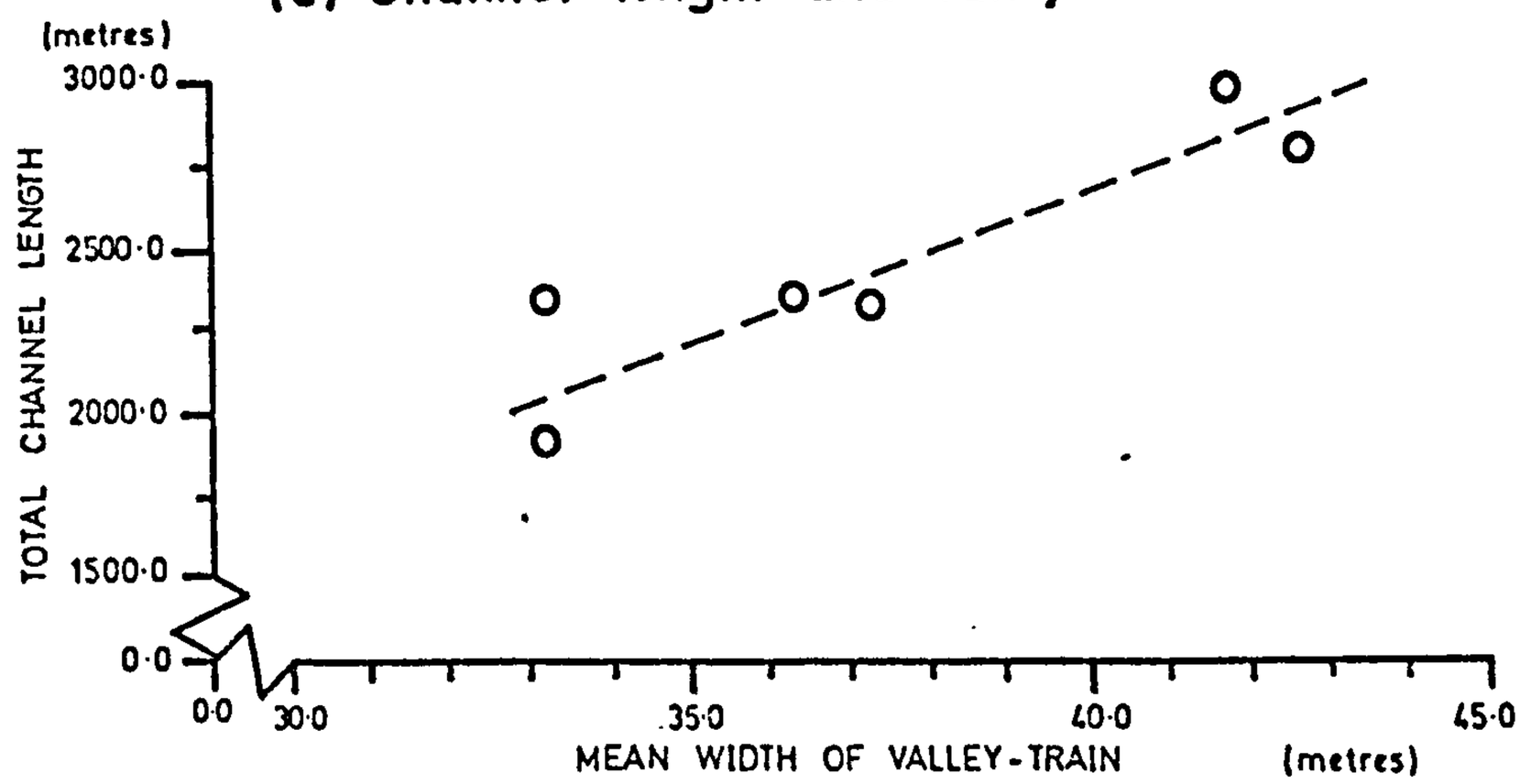
and $$L = 3(P+1) \qquad (5.4)$$

The resulting values are, however, difficult to interpret in physical terms.

FIGURE 5.4 SOME CHARACTERISTICS OF THE MELTWATER CHANNEL PATTERN ON THE BOSSONS VALLEY-TRAIN, FROM 1968 TO 1974







Church (1972) found this method unreliable because he was uncertain that he had measured all comparable channels, since the aerial photographs on which he based his measurements were taken at different stages of stream flow. Similarly, measurements at Bossons in July 1971 showed a drop of about 50 nodes from the April value, suggesting that seasonal flow variations affected the result by about 20 per cent. The scale of measurement also affects comparisons between different areas. All channels were surveyed at Bossons (Figure 5.5, page 239), whereas Church (1972) selected only the 'major' channels for his calculations. Over 300 nodes were recorded at Bossons in 1974 (Table 5.1, page 233), compared with a maximum of only 29 nodes from the much more extensive area of Lewis sandur in Baffin Island.

The increase in the number and total length of channels in the decreasing valley-train area of the Bossons glacier between 1968 and 1974 is directly reflected in the increase in the number of channel bifurcations (division points, P), junctions (nodes, N) and linking reaches (links, L) (Figure 5.6, page 240). The progressive increase in these topological indices reflects the increase in braiding as the meltwater streams have continued to aggrade the valley-train.

5.3.1.4. Bar morphometry

The size, shape, frequency and orientation of mid-channel bars (see Allen, 1968 in Collinson, 1970) closely reflects stream dimensions (Sundborg, 1956). The length of the bar is related to the length of stream elements between bifurcations and junctions; the width of the bar reflects the number of channels of particular widths in a given valley width; and the number of bars depends on the number of channels which have developed as a result of braiding mechanisms. An increase in mid-channel sedimentation to form bar cores, for example, will be associated with an increase of either the number of bars or the size of bars.

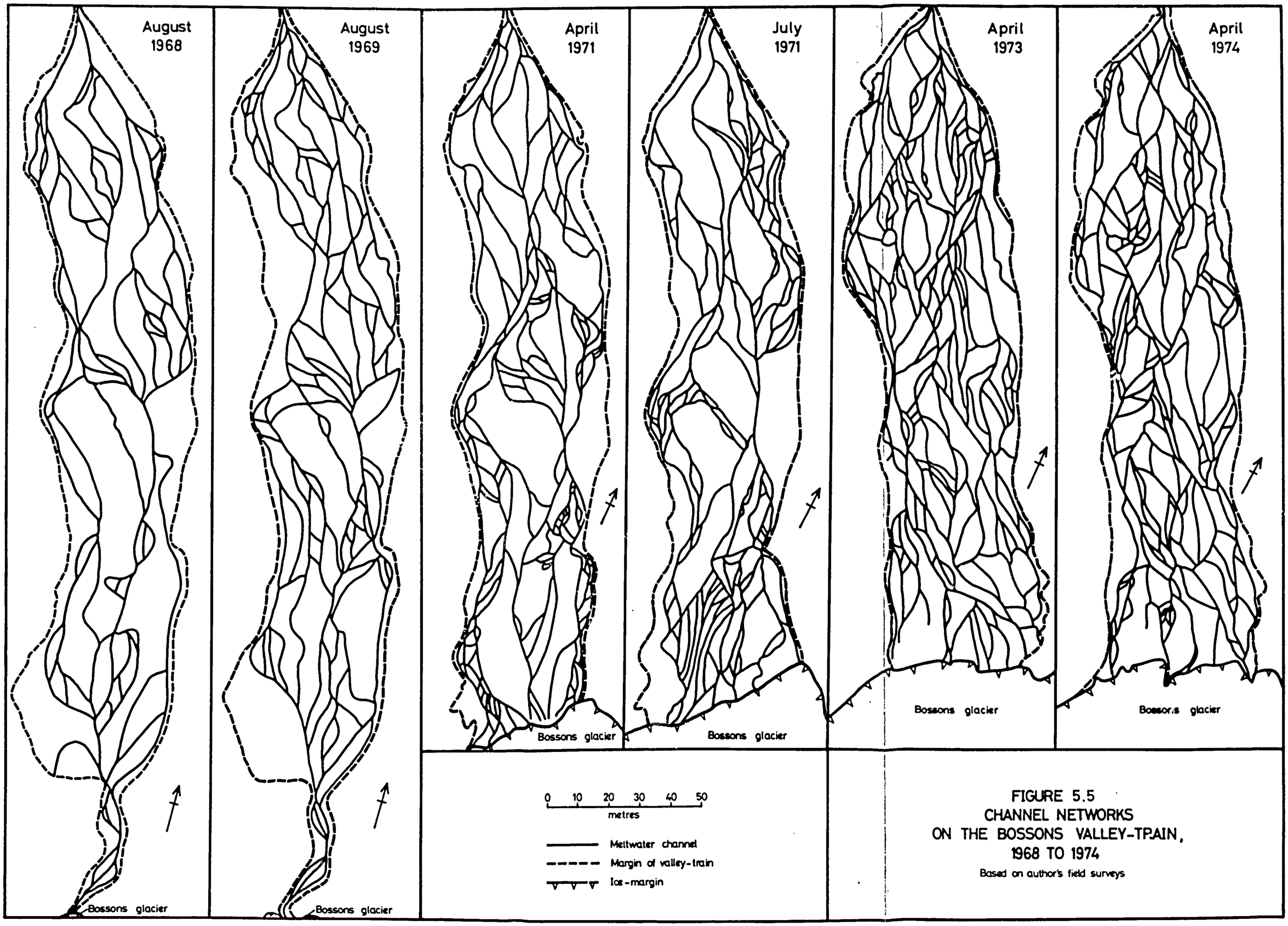


FIGURE 5.5
CHANNEL NETWORKS
ON THE BOSSONS VALLEY-TRAIN,
1968 TO 1974

Based on author's field surveys

Figure 5.6
Topological indices of channel pattern on the
Bossons valley-train, 1968 to 1974

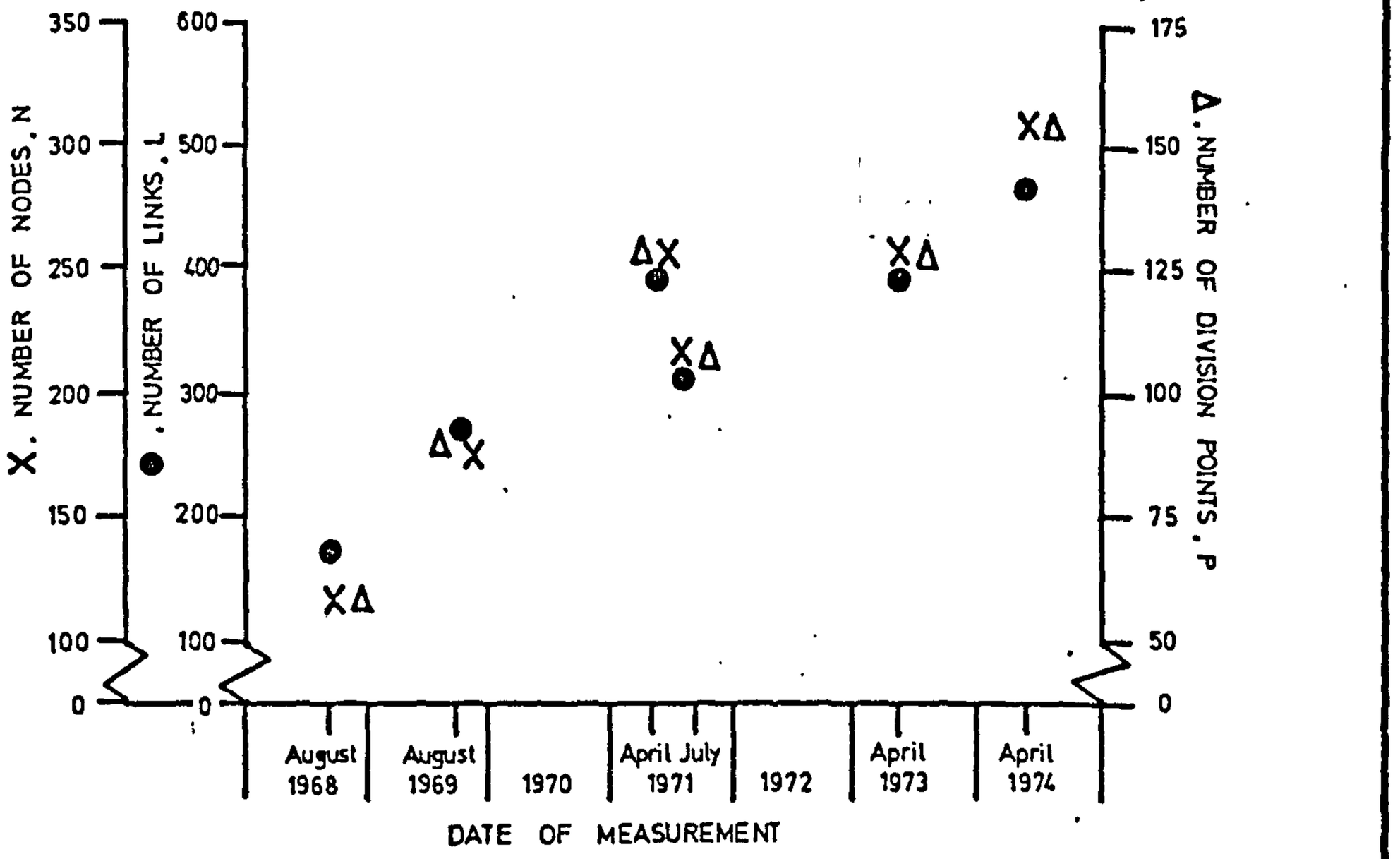
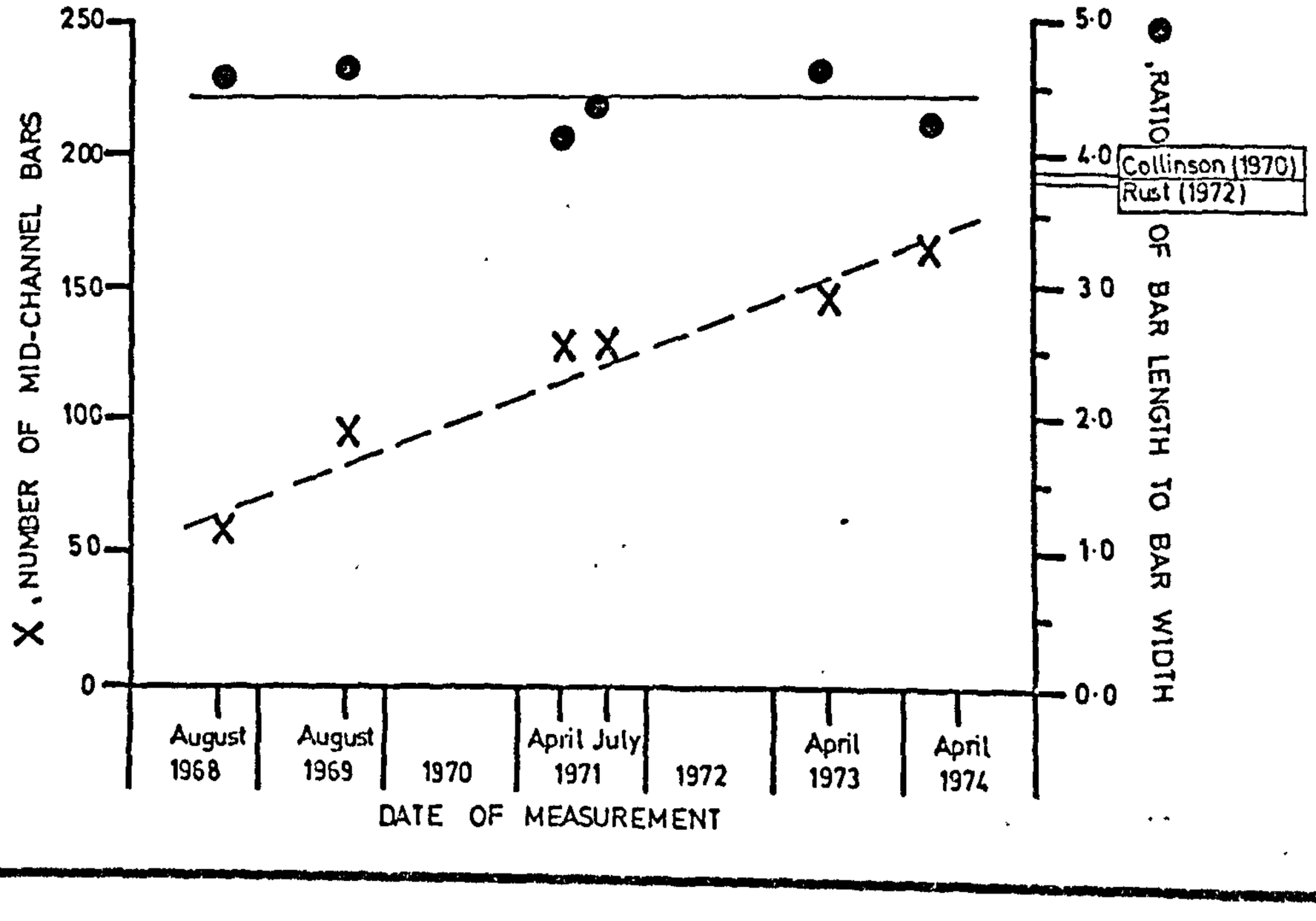


Figure 5.7
The frequency and shape of mid-channel bars
on the Bossons valley-train, 1968 to 1974



Measurements of bar dimensions on the Bossons valley-train between 1968 and 1974 show that the mid-channel bars have become progressively more numerous and much smaller (Table 5.1, page 233). Longitudinal mid-channel bars may be distinguished from transverse mid-channel bars, which are shorter, wider and composed of the more finely-grained sediments present in the downstream reaches of braided streams (Smith, 1970; Boothroyd, 1972; Rust, 1972). Longitudinal bars are predominant on the Bossons valley-train, although the frequency of both kinds of mid-channel bar has increased over the six-year period. The total number of bars increased by 276 per cent (from 58 to 160) (Figure 5.7, page 240), while bar length and bar width were reduced by 62 and 69 per cent respectively (bar length from 19.06m to 11.87m, and bar width from 4.18m to 2.88m). Reductions in bar width are directly associated with decreases in stream width over the same period. The 'shape' of the mid-channel bars, defined as the width-length ratio, has been maintained at each period of observation, since each dimension has been reduced by a similar amount (Figure 5.7, page 240). The average width-length ratios of mid-channel bars at Bossons remained between 4.07 and 4.69. These values compare closely with Rust's (1972) value of 3.90 for the bars on the Donjek valley-train, and with Collinson's (1970) values of 2.50 to 5.00 in the Tana River, Norway. The proportion of the valley-train occupied by bars has only varied by 12 per cent during the meltseason observation periods, ranging from 61 to 73 per cent, suggesting that stream aggradation has been accompanied not by the deposition of larger continuous masses of sediment but by more numerous pockets of sedimentation over the whole valley-train to form a more 'highly braided' system than formerly.

Brice (1964) introduced a 'braiding index' based on the length of bars, L_b, and the length of the reach measured halfway between the banks, L:

$$B.I. = 2\Sigma L_b/L \qquad (5.5)$$

B.I. values greater than 1.5 indicated a braided reach; values at Bossons ranged between 7.43 (in 1968) to 18.52 (in 1973). However, Brice admits that the index has no hydraulic or morphological significance, and it is felt that the measures already outlined above are simpler and more useful indicators of the degree of braiding (see also Collinson, 1970).

The orientation of mid-channel bars is determined by the directions of stream flow around the whole of the bar and its boundaries. The orientation of the long axis of bars may be used as a measure of the mean direction of stream flow. Bar orientation is not related to local flow deviations (Harms and Fahnestock, 1965; Williams, 1971), which cannot in any case be determined from the survey maps.

Mid-channel bars on the Bossons valley-train (Figure 5.8, page 243) mostly exhibit a distinct preferred orientation (Figure 5.9, page 244). Minor modes are apparent in the diagrams for August 1968 and July 1971 which represent summer discharge conditions when partial bar flooding and transverse flows across bars are more frequent.

The mean orientation of mid-channel bars altered progressively westwards between 1968 and 1974 (Figure 5.10, page 245), following three measured stages:

Mean orientation in (a) 1968 and 1969 was 342° (approx.),
in (b) 1971 was 330° (approx.),
and in (c) 1973 and 1974 was 328° (approx.).

It is possible that this westerly shift in bar orientation is caused by a change in valley-train orientation resulting from glacier advance. The Bossons glacier overran the southern part of the valley-train, mainly between stages (a) and (b) (see Chapter 2.4.3.2. and Chapter 3.1), filling in the western sector with moraine and debris fans, and the eastern sector with further outwash deposits. The proximal area of the valley-train has thus been extended eastwards,

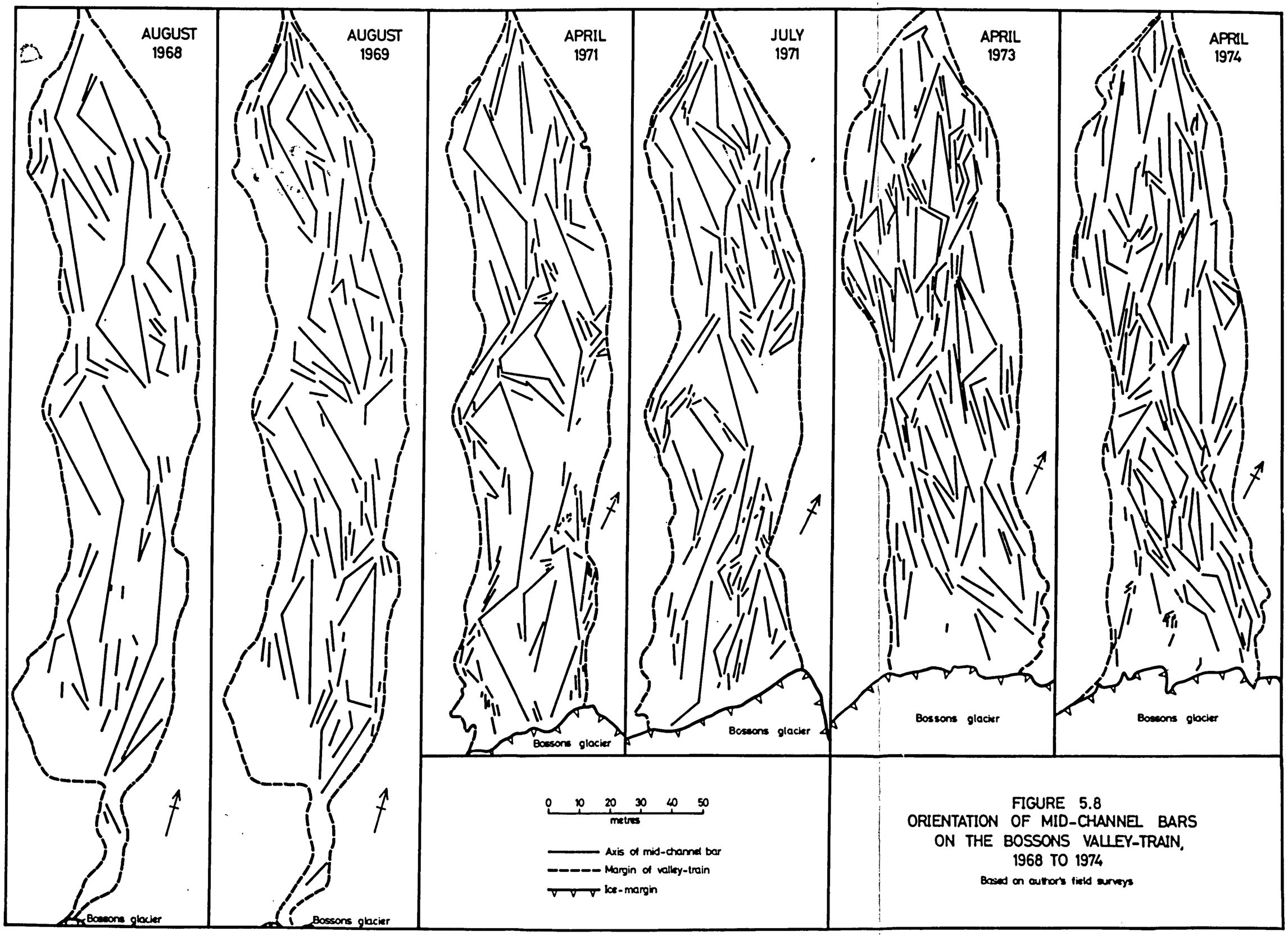


FIGURE 5.8
ORIENTATION OF MID-CHANNEL BARS
ON THE BOSSONS VALLEY-TRAIN,
1968 TO 1974

Based on author's field surveys

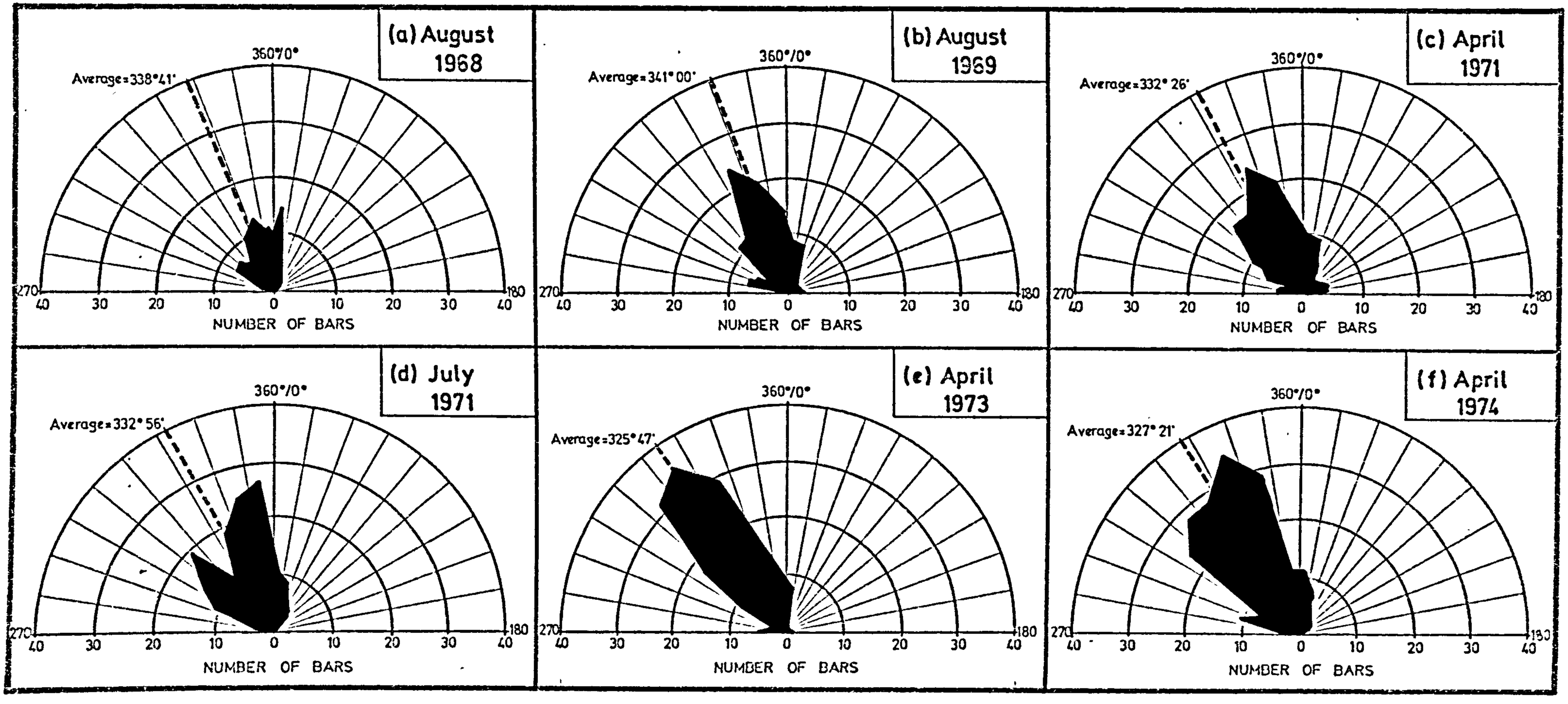


FIGURE 5.9 THE ORIENTATION OF MID-CHANNEL BARS ON THE BOSSONS VALLEY-TRAIN, 1968 TO 1974
Rose diagrams

Figure 5.10 The mean orientation of mid-channel bars on the Bossons valley-train, 1968 to 1974

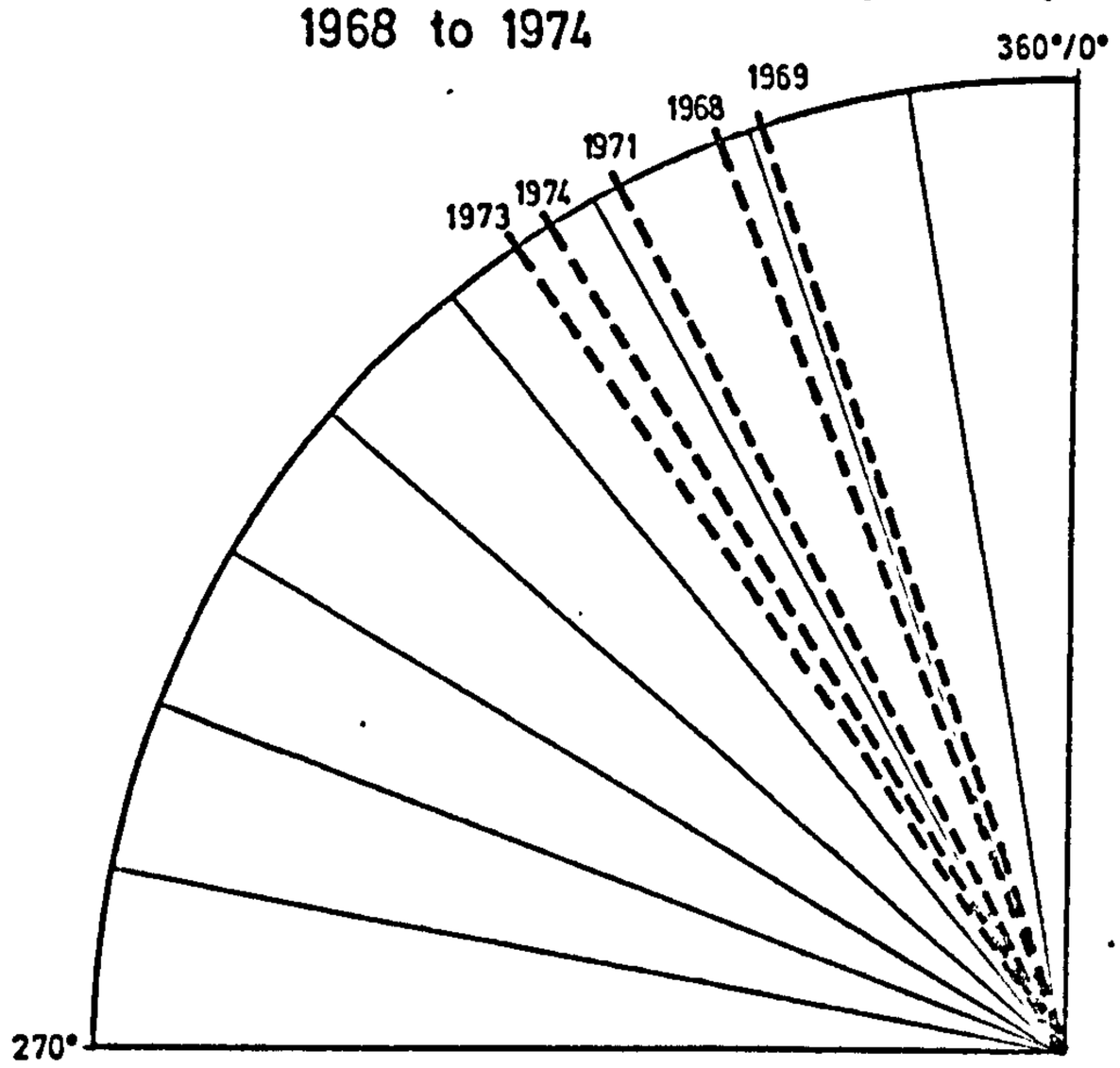
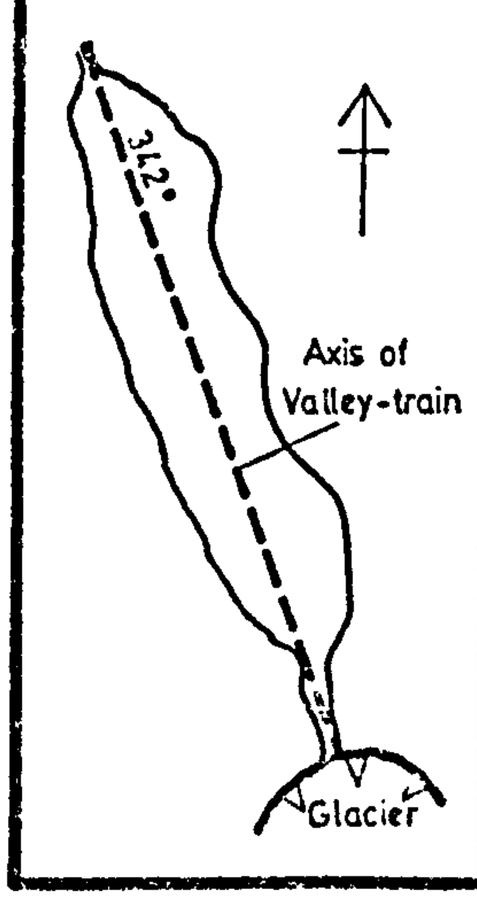
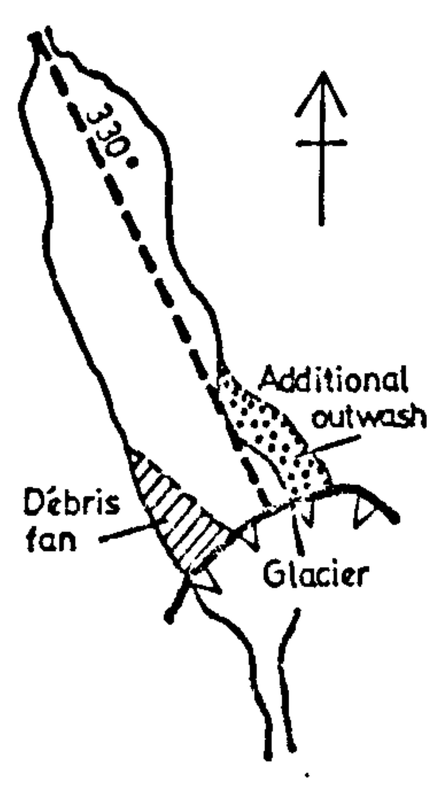
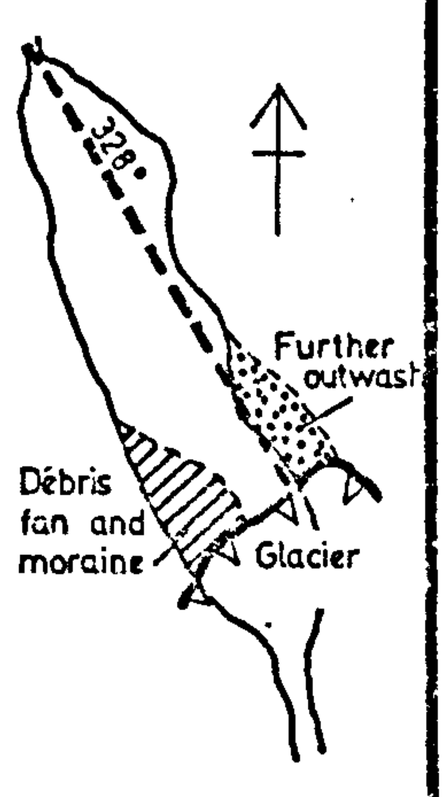


Figure 5.11
Diagrammatic explanation of changes of orientation of the Bossons valley-train, 1968 to 1974

(a) 1968 and 1969	(b) 1971	(c) 1973 and 1974
		

NOT TO SCALE

so that the valley-train orientation has shifted to include a more westerly component (Figure 5.11, page 245). The aggradation of the valley-train resulting from glacier advance has thus resulted in re-orientation of the valley-train axis, which is directly reflected in the changing mean orientation of mid-channel bars.

5.3.2. CONCLUSIONS; APPLICATIONS TO THE STUDY OF PALEO-OUTWASH DEPOSITS

The most marked effect of valley-train aggradation has been the progressive increase between 1968 and 1974 in channel and bar frequency, and a corresponding decrease in channel width and bar size, as a response by streams accommodating greater quantities of sediment. Fahnestock (1963) has shown that the number of channels is also related to discharge at the valley cross-section, and to distance from the ice-margin, as valley width and slope vary along the valley-train. Leopold, Wolman and Miller (1964) have demonstrated that in 'normal' fluvial basins, progressive drainage extension (by surface run-off, subsurface seepage, and headward erosion of streams) leads to a natural increase in the number of streams and total stream length. Similar processes may also operate in a braided network, where streams, fed by subsurface seepage, cut back into the bars to produce a highly dissected channel and bar system. The larger numbers of channels and bars would represent a natural progression, not necessarily related to changes in hydraulic and sediment load parameters. However, since aggradation has been known to occur, the numbers of channels and bars seem to be useful measures of changes in the degree of braiding which characterizes the stream network over a period of time.

The larger the number of bars and channels per unit area, the smaller the bars themselves, and the narrower the channels. Bar width at Bossons, for example, tended to be 1.5 to 3 times greater than channel width (mean = 2.4 times greater), but is generally 0.25 times smaller than bar

length. These approximate empirical relationships may prove useful in the analysis of paleo-outwash deposits where only a single parameter can be determined (see also Collinson, 1970). The length or width of a preserved and buried bar might give a clue to the former channel widths if no pairs of well-preserved channel boundaries are found to exist, e.g.

$$B_L = 4B_W = 4(1.5 \text{ to } 3.0)C_W \qquad (5.6)$$

where B_L is bar length, B_W is bar width and C_W is channel width ($B_L = 4B_W$; $B_W = (1.5 \text{ to } 3.0)C_W$; and $C_W = B_L(6.0 \text{ to } 12.0)^{-1}$). A possible bar deposit in the North Esk gravel pit, for example, extends for at least 160m (see Figure 5.26, page 294). The use of the relationships established for the Bossons outwash suggests that the width of the bar may be about 40m and the width of marginal channels may have reached a maximum of between 13 and 27m, according to flow stage. Further testing of these relationships is still required to establish their full range of reliability. Valid estimates of bar length, based on several paleo-depositional locations, might give an indication of the average distance between channel junction points, or nodes. Measurements from the Bossons channel pattern indicate that the nodes closely approximate to a log-normal distribution (see Appendix 1 and Figure A1.2). The nodal pattern as mapped in 1973 (Figure 5.12, page 248) also lies between a random and a scattered distribution, according to the calculated 'nearest neighbour statistic', R, which is 0.51 $\pm$ 0.04. Bar lengths are closely associated with the distances between nodes and also follow a log-normal distribution (Figure A1.2). It may be possible in future work (not presented here) to generate a log-normal pattern of nodal points, once $\bar{r}_a$, the mean distance between nodes, has been estimated from bar lengths. A series of channel systems could then be randomly interpolated between the nodes to give a picture of possible paleo-channel networks in the pro-glacial area. The patterns generated in this way would also have to take into account the probable degree of channel sinuosity and variability of local current direction and bar orientation.

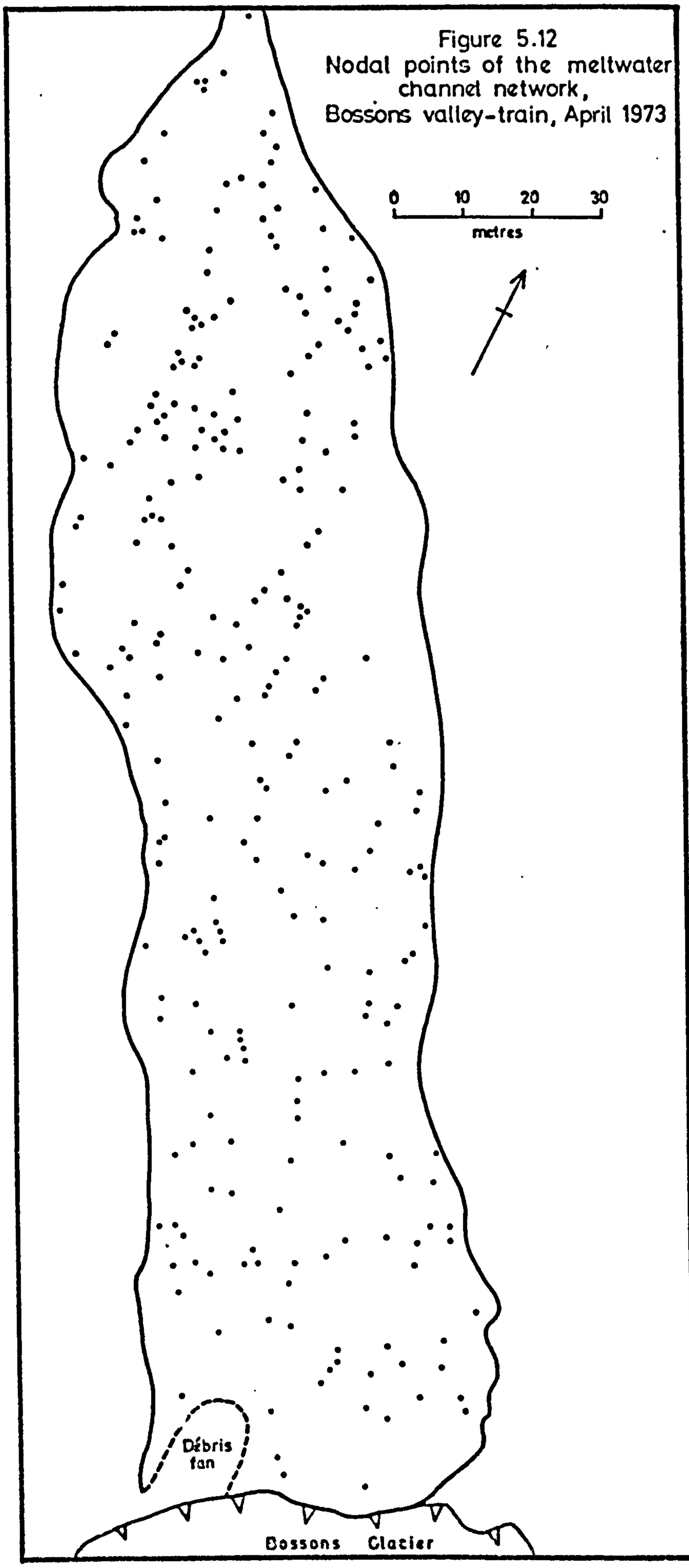


Figure 5.12
Nodal points of the meltwater channel network, Bossons valley-train, April 1973

Part of the predicted channel pattern could be checked if boreholes and/or exposures were concentrated in one area (see Spoljaric, 1974). The number of channels could then be estimated, so that, together with information on probable channel depths (see Section 5.4 and Chapter 7.6), velocities (see Section 5.5, Chapter 7.6 and Appendix 1) and regional slopes (Chapter 4.6), the total meltwater discharge of the former glacier could be calculated.

5.4. CHANNEL MORPHOLOGY

5.4.1. CHANNEL WIDTH AND DEPTH

The shape of a channel in cross-section reflects flow conditions, type and quantity of sediment load, and the gradient of the reach. Braided meltwater channels are typically wide and shallow, mainly because of the easily eroded, poorly sorted, unstable bank material (see Ackers, 1962, for calculation of the maximum height of stable banks), and the large quantities and rates of movement of sediment load (e.g. Lane, 1937; Leopold and Maddock, 1953; Sundborg, 1956). The ratio of channel width to channel depth is frequently used as an index of channel shape, although some workers have used measures of 'effective' width-depth ratios, based, for example, on depths of submerged bars (e.g. Brice, 1964) or on cross-sectional areas (e.g. Church, 1972). The width-depth ratios for the three stream sections at Bossons mostly ranged from 5 to 15 in proximal channels to 20 to 40 in the distal channel (Figure 5.13, page 250, and Appendix 1). Other workers (e.g. Krigström, 1962) have also noted downstream changes in channel shape from deep and narrow to broad and shallow. The downstream change in channel shape reflects not only a decrease in slope (see Chapter 3), but also a slight increase in the proportion of fine sediments ($\angle$ 0.063 mm) in the bank material, from 3.58 to 6.31 per cent. Schumm (1960a,b, 1961, 1968, 1971) has demonstrated that a high width-depth ratio is related to

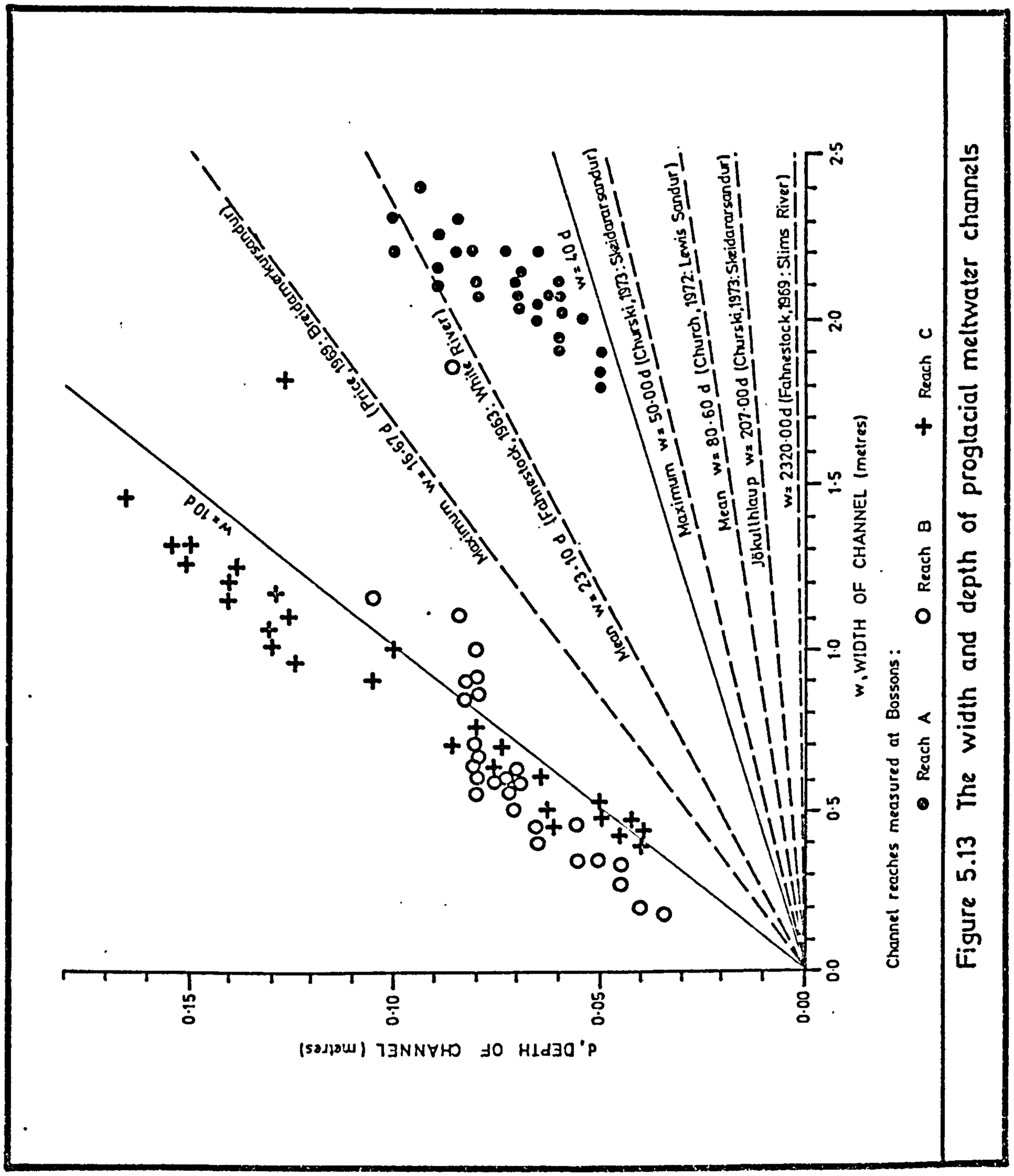


Figure 5.13 The width and depth of proglacial meltwater channels

small quantities of silt and clay in the bed and banks of a channel, regardless of stream discharge (see also Ackers, 1962). Wide, shallow channels contain little fine material but a large quantity of bedload material, partly derived from coarse, weak banks comprising only small amounts of cohesive silts and clays. Schumm's index of the 'percentage of silt and clay', M, is however, partly dependent on channel width (see also Figure 9 in Schumm, 1960a):

$$M = S_c w + S_B 2w/w + 2d \qquad (5.7)$$

where S_c and S_B are the percentages of silt and clay in the channel and bank respectively, w is channel width and d is channel depth. The wider the channel relative to its depth (i.e. high width-depth ratio, F), the greater will M be, given a constant percentage of silt and clay. Schumm's F/M relationship has, nevertheless, been widely used as a basis for defining the width-depth ratios of stream channels carrying different types of loads. Schumm has defined 'suspended load' streams as those in which the bedload forms less than 3 per cent of the total load, and which are, therefore, characterized by width-depth ratios of less than 10; 'mixed load' streams contain bedload amounts comprising 3 to 11 per cent of the total load and are characterized by width-depth ratios of 10 to 40; and 'bedload' streams are those in which the bedload forms more than 11 per cent of the total load and which are characterized by width-depth ratios of over 40 (Schumm, 1971). The proximal channels at Bossons appear to be, therefore, mainly 'suspended load' streams, where suspended load may represent over 95 per cent of the total sediment load. The stream in the distal channel, however, may be a 'mixed load' stream, in which suspended load represents about 90 per cent of the total sediment load. However, Schumm's classification is best restricted to alluvial channels in which less than 20 per cent of the sediment is of gravel size (Schumm, 1968). Furthermore, size analyses of the outwash sediments (Chapter 6.2) indicate

that nearly all the samples were deposited in bedload streams. Some proglacial streams may carry more than 90 per cent of their sediment load as bedload (e.g. Church, 1972).

Channel width and depth are thus partly dependent on the range of particle sizes present. Particle size is also important in determining the 'bed roughness' or friction presented by the channel perimeter to stream flow. The depth of flow in relation to the size of bed particles, for example, may influence the turbulence of flow, the rate of particle movement, the sizes of sediment carried in suspension, and the development of sediment accumulations on the channel floor, such as ripples and dunes or 'bar cores'. Some of the effects of shallow, coarse-bed channels on the methods of stream flow and sediment transport and on bedform structures are discussed in greater detail in sections 5.5 and 5.6 respectively. The relationships between stream width and depth, and velocity, and meltwater discharge in the three stream reaches at Bossons, are illustrated in Appendix 1.

5.4.2. APPLICATIONS TO THE STUDY OF PALEO-OUTWASH DEPOSITS

The proportion of silt and clay in a stream deposit can give clues as to the types of load transported by the stream, and to the likely range of width-depth ratios associated with the channels in which the sediments were formed. These relationships provide a method by which former channel depth, for example, may be predicted from knowledge of stream width and the percentage of fine material in the deposit (see Chapter 7.6). Sediment samples from the North Esk gravel pit indicate that the outwash deposits contain an average of only 0.275 per cent silt and clay (Chapter 6). The lack of sediments deposited from suspension suggests, therefore, that the North Esk meltwater streams were 'bedload' streams (see also Chapter 6.2). The width-depth ratio in this locality is likely to have been at least 40, while stream depths are likely to

have been small. The maximum width of 27m estimated earlier (section 5.3.2) for the channels adjacent to the buried bar in the North Esk gravel pit, would suggest that maximum depths were probably only about 0.7m (using d = w/40). Estimated values of former channel width, depth and thence cross-sectional area, together with measurements of particle sizes and paleo-slope, can prove valuable in determining many associated paleo-flow parameters, such as the nature of the sediment load, stream velocity and meltwater discharge (see Chapter 7.6, and Appendix 1).

5.5. MECHANISMS OF EROSION, TRANSPORTATION AND DEPOSITION IN A BRAIDED PROGLACIAL ENVIRONMENT

5.5.1. STREAM COMPETENCE

'Erosion' by a stream involves the picking up of particles from the perimeter of the stream channel for removal to a new position. The operation of this process depends on the critical stress required for particle motion to occur. The critical stress value can be expressed by the critical tractive force, τ_c, which reflects the 'competence' or ability of the stream to transport its sediment load, and is defined as

$$\tau_c = \gamma R_s S = \rho g d S \qquad (5.8)$$

where γ is the specific weight of the fluid, R_s is the hydraulic radius, S is the water-surface slope or energy gradient, ρ is the density of the fluid, g is the acceleration due to gravity and d is the depth of flow. The corresponding velocity required for movement of particles on the stream bed is the critical shear velocity, U_* where $U_* = (\tau_c / \rho)^{\frac{1}{2}}$. (5.9)

Critical flow conditions for particle motion are closely related to the size of particles, and many complex and simple functional equations have been derived to express this relationship (e.g. see O'Brien and Rindlaub, 1934; Hjulström, 1935; Shulits and Corfitzen, 1937; and Shields, 1936, in Graf, 1971). A useful simple formula for particles $\geq$ 6 mm has been introduced by Krey (see Graf, 1971), in which

$$\tau_c = 0.076 (\gamma_s - \gamma)D, \qquad (5.10)$$

where $(\gamma_s - \gamma)$ is the difference in the density of the sediment, γ_s, and of the fluid, γ, and D is the median particle diameter in metres. More complex relationships are widely employed by reference to their graphical equivalents in which other hydraulic variables such as shear stress and the degree of turbulence in the 'laminar sublayer' are also considered. The degree of turbulence affects the rate at which particles are hydrodynamically lifted from the channel floor through the development of eddies, and is expressed by the Reynolds number; rapidity or tranquillity of flow is expressed by the Froude number (see details in Appendix 2). Turbulence is increased by greater bed roughness, expressed by a friction coefficient such as Chezy's 'C' or Manning's 'n' (Appendix 2). Bed roughness is a function of the sediment load, of bed configurations (see Section 5.6) and of particle size (Leopold and Maddock, 1953). Relationships between critical flow conditions and particle sizes are theoretically more acceptable when they take account of flow conditions near the channel bed, such as Shields' widely accepted graphical function based on the critical tractive force (1936, in Graf 1971), rather than of average flow velocities, such as Hjulström's 'competency diagram' (Figure 18, p.298, in Hjulström, 1935)(see also Appendix 2).

Proglacial meltwater streams are characteristically highly turbulent and flow rapidly; Reynolds numbers, Froude numbers and friction coefficients are very high.

Calculation of the critical tractive force for the coarsest particles in a fluvial or fluvioglacial deposit can prove valuable in two ways. Firstly, one can determine the force of water on the stream bed which was required to erode, transport and subsequently deposit the particles composing the sediment. A comparison of the sediment sizes from different fluvioglacial outwash deposits can be related directly to different forces of meltwater derived from each glacier snout (see Chapter 7). Secondly, a large range of other flow and channel parameters may be deduced, once the

value of critical tractive force has been obtained. Channel depth, slope, mean velocity, degree of turbulence and type of flow can all be calculated, or checked by the use of the empirical formulae mentioned above, as well as from those discussed in Sections 5.2 and 5.3. These calculations may also provide information from which to determine the rate of transport of the sediments on the stream bed. Specific examples of the calculation of the critical tractive force and other flow parameters are reserved for Chapter 7, where differences in flow conditions are used in helping to account for differences in particle sizes present in each of the outwash deposits being studied, and in outwash deposits investigated by other workers. Discussion of basic stream flow concepts and formulae employed is presented in Appendix 2 and Chapter 7.6.

Some factors have not, however, been incorporated in the tractive force - particle size relationships, and must be considered when interpreting the comparative values. Firstly, the equations are related to homogeneous channel sediments, but if coarser particles are embedded in a sandy matrix, the coarser particles may be more difficult to lift than if all the particles were the same size. Similarly, the shape of the bed particles is important because flat pebbles are liable to become imbricated (inclined upstream, see section 5.6.3.1.(2)) and more resistant to hydrodynamic lifting. Elongated pebbles are more easily rolled downstream along the channel floor, with their long axes perpendicular to the direction of current flow (Lane and Carlson, 1954); well-rounded particles fall more quickly to the channel bed than do angular particles (Guy, 1970). Thirdly, the lack of material of a particular size may mean that high tractive forces still only transport fine grade material, if the coarser grades are absent. The absence of coarse sediments may increase the resistance of the sediment to erosion if it is sufficiently fine to be held

by cohesive or colloidal forces. An absence of fine sediments, however, may result in a more easily eroded coarse deposit, or less easily transported particles forming an "erosion pavement". Finally, interpretation of the tractive force and associated formulae should take into account the possibility of random, local effects such as random velocity fluctuations near the channel bed, local shielding by large particles or channel floor projections producing new flow patterns or flow discontinuities.

5.5.2. METHODS OF CHANNEL EROSION

The critical tractive force formulae indicate the major factors which determine the ease and rapidity of particle erosion in a stream channel. The rate of erosion depends, in particular, on stream velocity, and rates of change of stream velocity, channel depth and slope, bed roughness, particle size and sorting, particle shape and packing, and the total quantity of the stream load. Fluvioglacial outwash deposits are highly susceptible to erosion, largely as a result of factors relating to stream régime and to the nature of the channel sediment:

(1) Stream régime Rapid and frequent changes in stream velocity and stream level provide greater opportunities for picking up material during periods of increasing velocity and increasing critical tractive force. Relatively frequent flood events during the meltseason mean that erosion can be effective over most of the exposed outwash surface. Repeated flooding reduces the stability of the substratum and often prevents colonisation by stabilizing plants, so that instability can be maintained over long periods of time.

(2) Nature of the channel sediment The outwash deposits are generally poorly sorted, heterogeneous materials, including, therefore, a wide range of particle sizes at each locality. Débris is consequently available for erosion at almost each value of critical tractive flow in the stream. The fine sands can be removed at lower

stream velocities leaving a coarser lag deposit, or removing the matrix from the channel bank which subsequently collapses.

The methods of channel erosion in the unconsolidated, heterogeneous sediments which typify outwash deposits have been widely discussed in the literature. Channel erosion may be grouped into erosion which proceeds laterally into the channel banks, and that which proceeds vertically, into the channel floor.

5.5.2.1. Bank erosion

Lateral channel erosion takes place in unconsolidated deposits as a means of hydraulic readjustment by the stream to changing flow conditions. An increase of channel width by bank erosion can result from the stream's attempt at accommodating an increase in the local discharge of meltwater and/or of sediment load (Schumm, 1961,1971). Channel banks characterized by a low threshold of erosion are more easily extended to cope with new flow and sediment conditions than is the channel floor, especially in an aggrading proglacial system.

Some of the mechanisms of bank erosion have already been mentioned, and include saturation by rising water levels, selective removal of the sediment matrix and some coarser particles, leading to undercutting and slumping (Figure 5.14, page 258). Bank erosion is most effective on the outer bends of streams where the greater force of water allows most of the slumped debris to be removed, thus exposing a further thickness of bank to stream erosion processes. Continual bank recession causes the channel to migrate laterally across the outwash surface (see also Nordseth, 1973b). Many braided channel networks are characterized by frequent shifts in channel position, usually resulting from progressive bank erosion.

5.5.2.2. Channel bed scour

Vertical scour of the channel bed also represents a change in channel shape in response to alterations in stream discharge or sediment load. Bed scour appears to

Figure 5.14
Lateral bank erosion on the Bossons valley-train

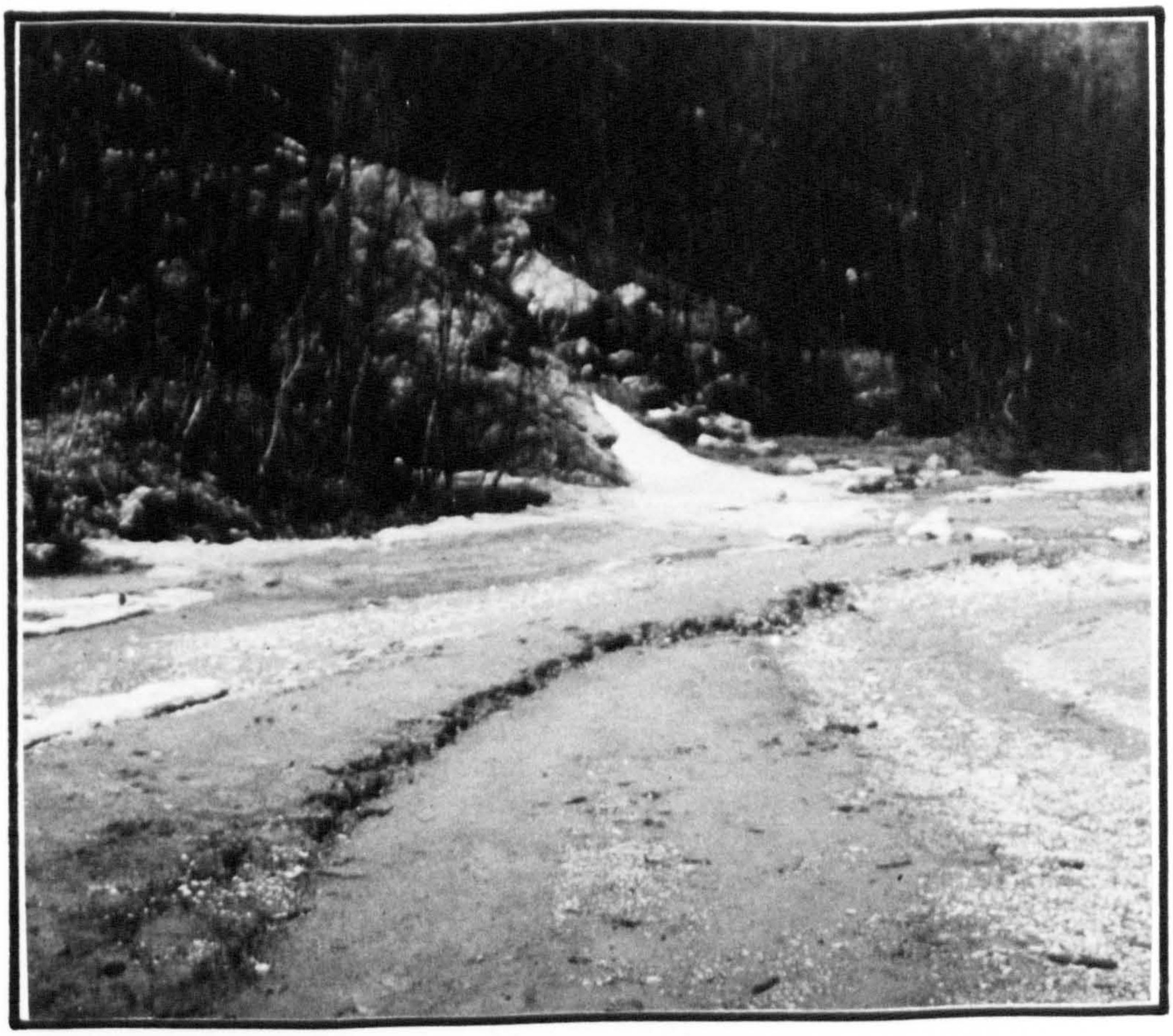
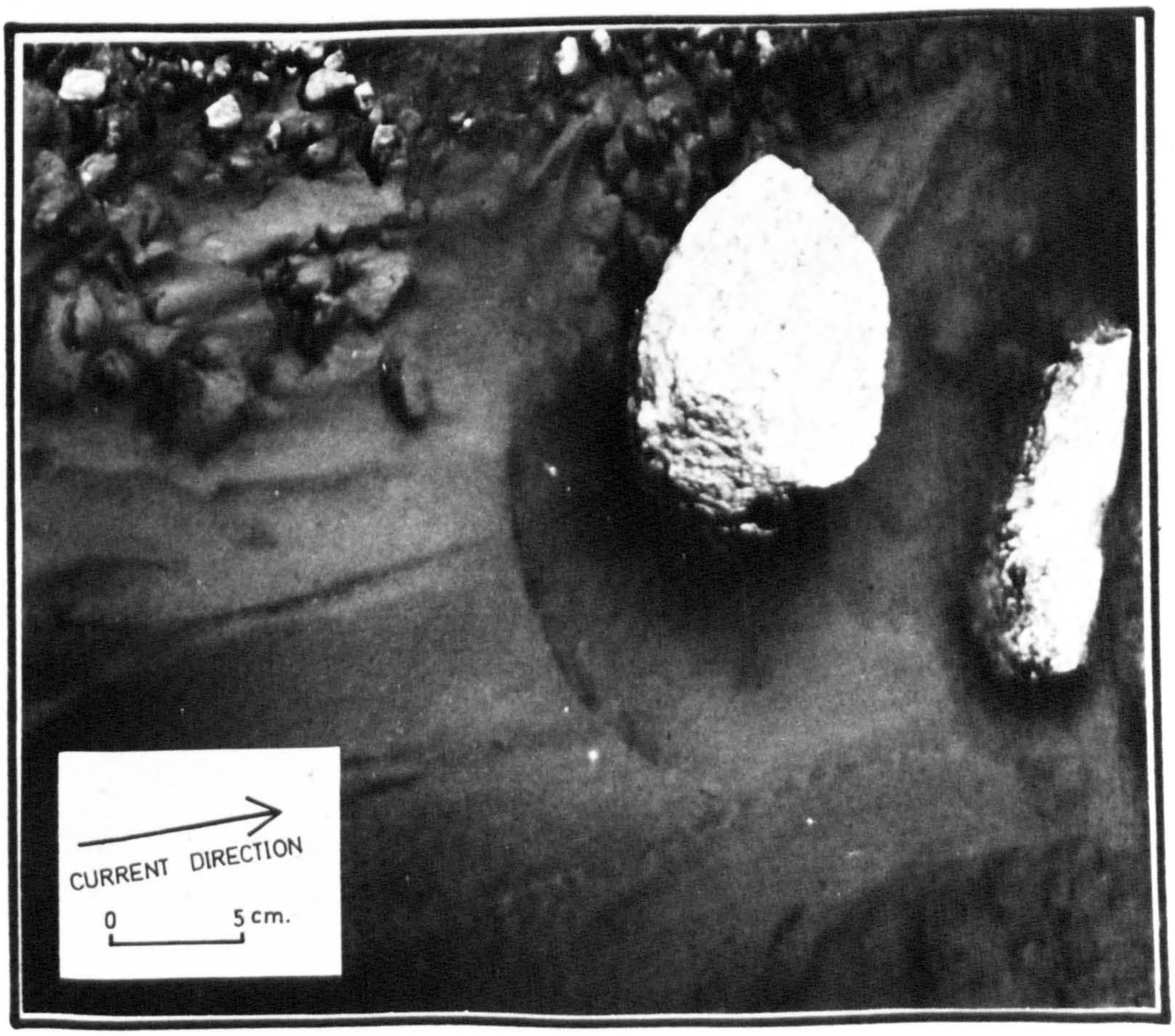


Figure 5.15
Current crescent scour upstream of rock fragment,
Bossons valley-train



occur when a change in stream velocity is accompanied by a local deficit of sediment load. Excess stream energy available for local downcutting becomes available during either an increase or a decrease of discharge,as long as a lesser increase (or greater decrease respectively) of sediment load also takes place. Leopold, Wolman and Miller (1964) recorded bed scour during periods of increasing stream velocity, while Fahnestock (1963) observed channel scour during periods of low velocities or decreases in velocity. The occurrence of bed scour, therefore, depends not only on velocity or velocity changes, but also on the debris supply. The nature of the bed material also plays an important role in determining the ease with which particles may be removed from increasing depths on the channel floor. A coarse 'lag' deposit or 'erosion pavement' increases resistance to erosion, encouraging lateral rather than vertical scour. Scour is usually a relatively localized, short-term channel phenomenon (Sundborg, 1956), most effective in areas where flow converges and becomes concentrated before diverging again. Water flowing down the downstream face of a partially submerged channel bar, for example, forms a 'flow chute'. The chute is progressively scoured and works its way upstream through the bar (Church, 1972). Scours may, alternatively, be quickly infilled by sediments avalanching down the bar slope (Harms et al., 1963). Scour can also take place upstream of a coarse particle, boulder or other channel bed obstruction (e.g. Fahnestock and Haushild, 1962; Moss, 1963; Collinson, 1970). Flow is concentrated upstream of a boulder where scour takes place to form a 'current crescent' (e.g. Williams and Rust, 1969; Rust, 1972) (see Figure 5.15, page 258). Stream flow divides round the obstacle, and deposition takes place in the shallower, narrower channels downstream (McDonald and Banerjee, 1971). Scoured hollows also develop downstream from the re-entrants in the crest of dunes and ripples (e.g. Williams, 1968).

Mid-channel scour is, however, the most prevalent form of channel deepening (Fahnestock, 1963), and occurs

during peak flows when channel width and depth are increased to cope with greater stream velocities and sediment loads. These scoured channels are usually filled with sediment during the declining stages of the flood as tractive force decreases. Deposition reduces channel depth to produce "scour-and-fill" structures (Section 5.5) which are frequently, therefore, the product of a single flood event.

Channel scour can destroy the uppermost levels of any underlying sedimentary structures. Distinct 'erosion surfaces' can often be observed in paleo-outwash deposits which indicate where a new flow régime has brought in a new series of sediments into the channel. 'Sheet erosion' can also occur during periods of overbank flow. The whole submerged outwash surface may become mobile, and huge quantities of sediment may be moved in a single flood. The base of such deposits may also indicate initial bed scour (Figure 5.16, page 261).

5.5.3. TRANSPORT OF SEDIMENT LOADS IN FLUVIOGLACIAL STREAMS

A particle, or mineral element, may become part of the dissolved, suspended or bed load of the stream. The type of load depends largely on sediment parameters (e.g. composition, degree of weathering, density, size, shape, packing) and hydraulic parameters (e.g. temperature, viscosity, velocity, turbulence of meltwater). The differentiation between loads mainly depends on the rate at which particles are able to fall to the bed of the stream. Particles $\leq$0.1 mm in diameter face resistance from the viscous fluid and settle at rates according to Stokes' Law, which states that the fall velocity, v, is largely proportional to the square of the particle diameter, D (e.g. Hjulström, 1932), such that

$$v = C\left(\frac{D}{2}\right)^2 \qquad (5.11)$$

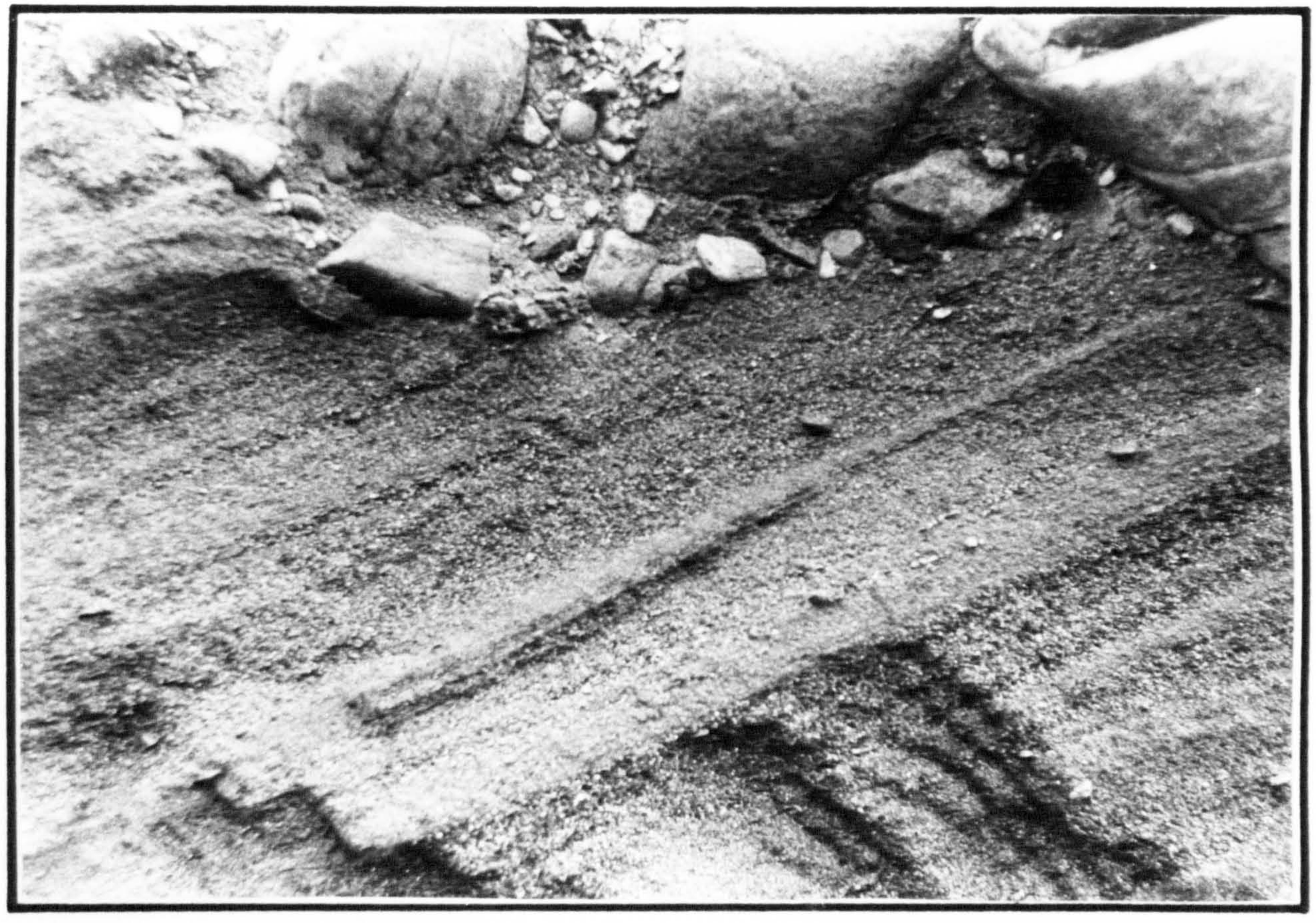


Figure 5.16 Erosion surface in paleo-outwash deposits, South Esk gravel pit

where C is a constant[1]. The settling velocity of particles whose diameter exceeds about 1.0 mm is independent of fluid viscosity, and is, instead, dominated by the effects of particle surface area, sphericity and turbulence, such that

$$v = C\left(\frac{D}{2}\right)^{\frac{1}{2}} \qquad (5.13)$$

which is the impact law of settling (Fuller, 1961; Krumbein and Sloss, 1963; Griffiths, 1967; Allen, 1970a). The increase in particle size results in turbulence effects (see Sundborg, 1956), which are associated with large Reynolds numbers. The mean of Stokes' Law and the impact law applies to particles between 0.1 and 1.0 mm in diameter.

Erosion and run-off within the drainage basin determine the sediment loads and discharge of water entering the stream (see Leopold and Maddock, 1953). The type of sediment load subsequently carried by the stream depends on relationships between the size and shape of the particles, morphology and roughness of the channel, and the velocity and turbulence of the stream. Each of these parameters may vary continuously at one section. The amount of sediment load of each type is therefore also variable at each section over time, and sediments in suspension may quickly become part of the bedload, and vice versa. All three types of sediment load (in solution, in suspension and bed load) are found in proglacial streams. Schumm (1971) has classified braided channel systems as 'bedload channels', in which more than 11 per cent of the total load is transported along the stream bed.

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1 $$C = \frac{2}{9}\,\frac{(\gamma_s - \gamma_f)g}{\eta} \qquad (5.12)$$

where v is the settling velocity (cm/s), γ_s is the density of the sediment (taken as 2.65 g/cm^3), γ_f is the density of the fluid (taken as 1.00 g/cm^3), g is the acceleration of gravity (taken as 980 cm/S^2), and η is the viscosity of the fluid (taken as 0.0152 poises, at 5°C or 0.0179 poises at 0°C). Using the values given here, C = 2.364 x 10^4 at 5°C and atmospheric pressure (Griffiths, 1967).

5.5.3.1. Sediments in solution in proglacial streams

(1) Solution processes in cold environments

Glacial meltwaters characteristically contain high quantities of mineral ions in solution (Thorarinsson, 1939; Corbel, 1959; Keller and Reesman, 1963; Vivian, 1970c). The solution process has been considered by these authors to be a significant agent of chemical weathering even in cold environments. Other workers have also cited evidence of active chemical weathering processes operating in Arctic and proglacial regions. Bird (1967) observed relatively rapid rates of leaching where organic acids were present, and many workers have recorded the oxidation of iron-rich rocks and minerals in periglacial conditions (e.g. Hill and Tedrow, 1961; Bird, 1967; Church, 1972). Dahl (1966) states that chemical weathering can also take place in frozen bedrock. These workers claim that the most important process is the solution of mineral elements by waters charged with CO_2 (carbon dioxide). Williams (1949) found that the solubility of CO_2 is greatest at temperatures close to freezing point (see also McCann et al., 1972). The CO_2 is derived from the turbulent action of streams at the water-air interface, from certain other chemical reactions such as the solution of bicarbonates (Vivian, 1970c), from the compaction of ice (Ek, in Cogley, 1972), and from the freezing and subsequent melting of snow (Williams, 1949). Snow is initially quite acid (pH 4.4 to 7.0; Clement and Vandour, 1967). Snow and meltwater which freeze release CO_2, while melting provides water which absorbs the CO_2 to form carbonic acid ($H_2O + CO_2 \rightarrow H_2CO_3$) which is an active solution agent (Keller and Reesman, 1963; Clement and Vandour, 1967; Pitty, 1971).

Williams' results did not take account of any other weathering processes than O_2 and CO_2, he made no reference to the nature of the bedrock around the snow patches, and he obtained only three samples for analysis (see Smith, 1972). More recently, other workers have recorded much lower rates of solution weathering than suggested by Williams (1949) and

Corbel (1959), in particular. Church (1972), for example, described very low rates of solution in Baffin Island, being equivalent to surface lowering of between 0.69 and 1.52 mm/1000 years. Similarly, McCann et al. (1972) found that the rates of solution are much lower in colder regions than in the warmer more humid regions, although CO_2 solubility is higher (see also Peltier, 1950). The total amounts of material in solution in meltwaters are, therefore, not necessarily greater than in lower latitude streams. Smith (1972) also recorded low rates of solution of limestone in Arctic regions, producing a surface lowering of 3 mm/1000 years. This recent evidence suggests that solution weathering in colder regions is of relatively minor importance.

The dissolved meltwater load, nevertheless, represents the major chemical weathering process producing mineral ions, such as Si, Ca, Mg, K and Na, for removal from the glaciated basin through the action of carbonic acid. Carbonic acid also becomes effective in the hydrolysis and disintegration of mafic rocks (e.g. Keller and Reesman, 1963; Pitty, 1971; see also Chapter 6.5). The soluble stream load is also present throughout the year, including periods when discharge is too low for the other types of sediment load to be moved (e.g. McCann et al., 1972). Vivian (1970c; Vivian and Zumstein, 1973) recorded an excess of dissolved load over suspended load from November to April for several Mont Blanc glacial meltwaters, and the reverse situation during the meltseason, from May to October when the ion concentrations are diluted in larger volumes of meltwater (see Cogley, 1972). Vivian concluded that chemical weathering might be the only important weathering process to act during the winter, although it is still on a small scale compared with other forms of weathering and removal of rock debris.

(2) Rates of solution in the Bossons glacier basin

The total quantities of dissolved materials removed by meltwaters from a glaciated basin can be determined by the calculation of a sediment-rating curve (e.g. Douglas, 1964; Church, 1972). The concentration of dissolved sediments is usually closely correlated with meltwater discharge, and once the curve of this relationship, and the total annual discharge, are known, the total amount of sediment removed in solution can be calculated. Corbel (1959) employed a different method, in which his calculations of volumes of sediment removal, A, are based on two run-off parameters: (1) the depth of water in the stream (or the difference between total precipitation and total evaporation, E, when no discharge data are available) and (2) the total discharge of sediments in solution (or in suspension), T, such that

$$A = 4ET \times 10^{-2} \qquad (5.14)$$

Corbel then estimated that a minimum of $200m^3/km^2/yr$ of sediments in solution are removed from the basin of the Bossons glacier. This value represents an average surface lowering of 200 mm/1000 years, which seems unacceptably high (compare Church (1972), Smith (1972)). Smith (1972) derived a more accurate formula which takes account of the rock density, D, and the amounts of $CaCO_3$ (T_c) and $MgCO_3$ (T_m) in the stream:

$$X = E\ (T_c + T_m)/10D \qquad (5.15)$$

where E is equivalent to Corbel's E where no discharge data are available, and X is the overall rate of solution lowering of the surface in mm/1000 years. Vivian (1970c; Vivian and Zumstein, 1973) quote a maximum content of 10 mg/l of minerals in solution in the meltwaters of the Argentière glacier, which would produce an average surface lowering of 6.23 mm/1000 years. Several measurements made by the author (Maizels, 1970) of $CaCO_3$ content in the meltwater of the Bossons glacier produced a comparable value of 7.48 mm/1000 years, while results from a simple formula derived by the

author (see below) suggested an average rate of surface lowering of 9.00 mm/1000 years:

$$X = \frac{(q_s)(Q_w)}{(A)(10^6)} \ (m^3/km^2/yr = mm/yr) \qquad (5.16)$$

where X is the amount of sediment removed in solution, q_s is the mean discharge of solution sediments, Q_w is the total annual meltwater discharge, and A is the glaciated area (see Table 5.2, page 274).

The quantity of sediment removed in solution cannot yet be determined for paleo-outwash basins. The amounts of dissolved material is not of any importance in the analysis of paleo-outwash deposits, since they were not deposited, but were transported directly to the sea. The suspended and bedload sediments, however, comprise the total sediment load of outwash deposits, and measurement of their respective proportions and sizes can therefore prove most valuable.

5.5.3.2. Sediments in suspension in proglacial streams

(1) Movement of particles in suspension

The suspended sediment load comprises particles whose motion is transitional between that of particles in continuous solution, which flow at the same velocity as the water, and that of the bedload, which moves intermittently along the channel floor. Sediments are brought into suspension whenever the turbulent forces inducing an upward velocity component (V_u) exceed the settling velocity of the particles (V_s), or when

$$V_u > 0.17 \ (V_s D)^{0.46} \qquad (5.17)$$

(after Krey and Jasmund, in Graf, 1971). Particles are lifted up into the stream flow either by hydrodynamic forces of uplift around the particle or by saltation (following collision with bed particles). The grains are then brought into motion. The finest particles may never be allowed to settle and are carried within the stream at all times. This continuous suspension load is known as the

'wash load'. Other particles may be carried in the stream, following oscillating vertical paths downstream before falling once more to the bed when turbulence temporarily decreases. The length of each 'saltation' and the height to which each grain is lifted depends on the grain parameters and hydraulic conditions at the time. This 'saltation' load is really transitional between the suspended load and the bedload (Section 5.5.3.3.). The concentration of sediments in suspension decreases from the 'bed layer' up to the stream surface, although the vertical differential is least in highly turbulent streams (Guy, 1970). The size of particles in suspension may also vary, therefore, according to the depth in the stream.

(2) Variability of suspended sediment concentrations

The quantities of suspended sediments in proglacial streams are extremely high when compared with quantities in streams from non-glacial basins (e.g. Lanser, 1958; Corbel, 1959; Østrem et al., 1967). The large volumes of suspended sediment in meltwaters are generally considered to reflect the great effectiveness of glacial erosion in the basin (e.g. Thorarinsson, 1939; Embleton and King, 1968). Many workers have attempted to estimate the annual amount of glacial erosion from the annual quantities of sediment removed in suspension by meltwaters from a glacier basin (e.g. Thorarinsson, 1939; Corbel, 1959; Church, 1972; Østrem and Ziegler, 1973).

Several workers regard the accuracy of such measurements as doubtful. Church (1972) states that the suspended sediments are not necessarily derived from recent glacier erosion. Most of the channel debris may have been eroded by Pleistocene or "Little Ice Age" glaciers, and deposited as moraine and fluvioglacial sediments which are only being reworked during the present-day.

Measurements of suspended sediment concentrations are also subject to variations due to changes in temperature, and therefore, the viscosity,of the water (e.g. Keller, 1916; Hjulström, 1932; Borland, 1961; Nordin and Beverage, 1965; Guy, 1970). The particle settling velocity decreases with

decreasing temperatures, so that the fine fractions in particular, remain in suspension longer in glacial meltwaters (temperatures <2 to 5°, see Vivian, 1970c and Maizels, 1970) than in warmer streams. Any increase in temperature, following a summer rain storm, for example, can act to accelerate the rate of deposition of material formerly carried in suspension (see data in Guy, 1970).

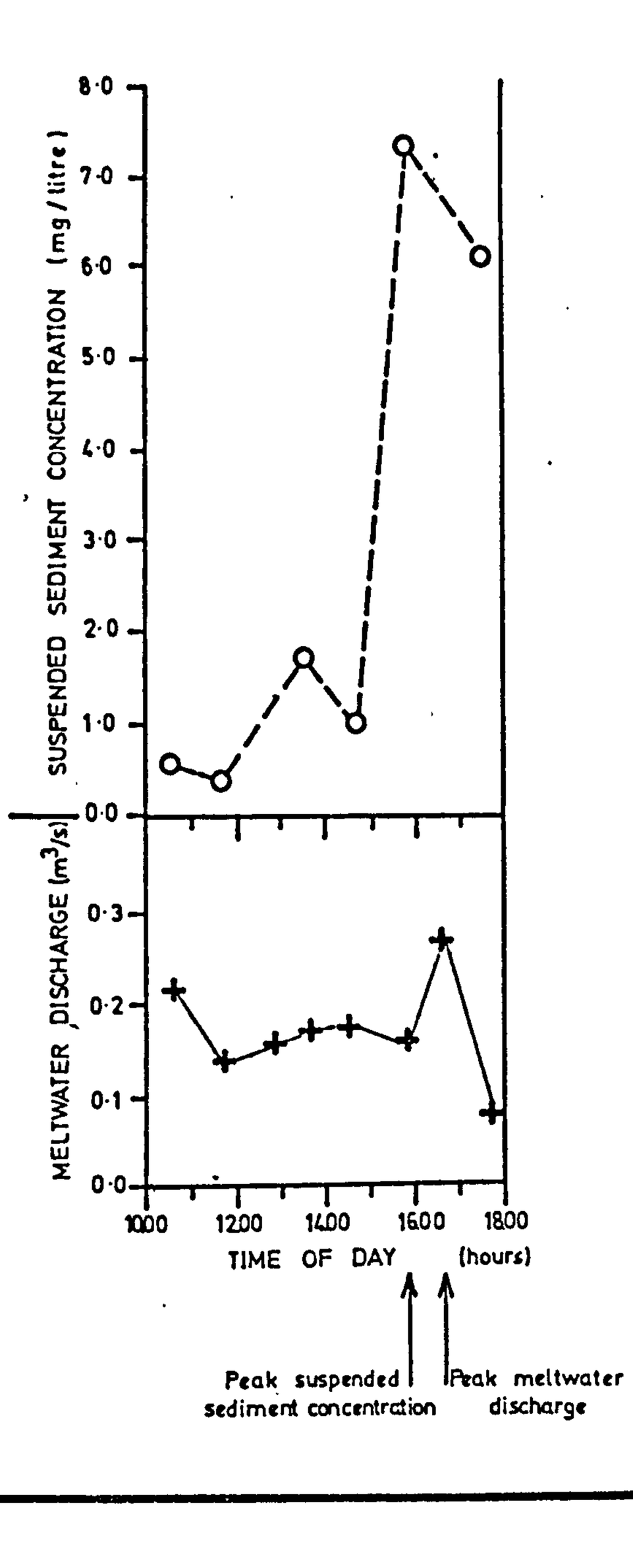
Direct measurements of suspended sediment loads, for calculation of rates of erosion, should ideally be carried out over a period of years, rather than over a single meltseason, because of the extreme variability of concentrations of suspended sediment in meltwater streams. Suspended sediment concentrations tend to increase with increasing distance at which the sampling site is located downstream from the ice-snout (also noted by Doeglas, 1962; Broscoe, 1972). Samples of meltwater were collected from Bossons outwash reaches A, B and C (Figure 2.5, page 62). Meltwaters from reach B (closest to the ice-margin) in July 1971 contained an average of 4522 mg/1, compared with 27620.61 mg/1 at reach A (over 200m from the ice-margin). Higher concentrations at downstream sites suggests that most of the suspended sediments may in fact be derived from the outwash deposits themselves, particularly during non-flood flow conditions. Meltwater at reach C (close to the opposite side of the ice-margin from reach B) flowed over subglacial and ice-marginal moraine, rather than directly over the ice, so that suspended sediment concentrations averaged 24714 mg/1 over the same period.

Variations in suspended sediment concentrations also occur over time (e.g. McCann et al., 1972). Seasonal, diurnal and random variations occur, but are only partially related to similar variations in meltwater discharge. Average suspended sediment concentration at reach A at Bossons, for example, was about 9 times greater in July than in April (27620.61 mg/1 compared with 3072.23 mg/1), while meltwater discharge increased by only about 5 times (0.5034m^3/sec. compared with 0.0985m^3/sec.).

Østrem et al. (1967) found that suspended sediment contents were higher in spring than at the end of the meltseason, for the same meltwater discharge. The higher concentrations in spring reflect the 'washing out' of fine sediments which accumulated during the lower discharges at the close of the previous meltseason, and during the winter. The cumulative effect of the fine deposits diminishes throughout the meltseason until the suspended sediment content represents only the debris made available to stream erosion during that season (see also Leopold and Maddock, 1953).

Many workers have measured the daily variation in run-off and suspended sediment content of proglacial streams (e.g. Fahnestock, 1963; Østrem et al., 1967; Liestøl, 1967; Church, 1972; Østrem and Ziegler, 1972). Their results were similar to those found at Bossons, for the maximum sediment content generally occurs up to 2 to 3 hours before the daily peak discharge (Figure 5.17; see also Nordseth, 1973b). The suspended sediment concentration is thus related to proportionate increases in discharge, rather than to actual discharge amounts. Sudden storm or flood events often provide sufficient increases in meltwater discharge to transport far greater amounts of sediment in suspension than can moderate flows. Østrem et al. (1967) reported that over 60 per cent of the total suspended sediment removed p.a. were removed in a single day, and other workers (e.g. Douglas, 1964; Church, 1972) have also measured extremely high quantities of suspended sediment during peak flows. Variability in suspended sediment concentrations also depends on rates of sediment yield from within the basin. Glacial erosion provides one source of sediment, but surface run-off from unvegetated areas, especially after periods of rain, can increase the sediment contents of streams.

Figure 5.17
Suspended sediment concentration
and meltwater discharge
Reach C, Bossons valley-train,
July 24 1971, 1000 to 1800 hours



(3) Measurements of suspended sediment concentration

(a) Methods. Measurement of the amounts of sediment in suspension should therefore be carried out over as long a period as possible, and application to estimates of rates of erosion should be considered with some caution. The total quantity of sediments removed from a glaciated basin may be calculated in several ways, although most calculations must,

at this stage, be only approximate, with few estimates of the possible degree of error.

The most accurate measurements of suspended sediment are based on the collection of water samples from a known depth in the stream. The vertical gradient of sediment concentration, and thence the total suspended load at the measured discharge, can be determined by incorporating the measurements in a sediment diffusion equation such as that proposed by Rouse (1937, in Graf, 1971; and Blatt et al., 1972).[1] The total suspended load in the stream at each discharge value can sometimes be successfully plotted as a sediment rating curve, which provides a fairly close predictive, logarithmic relationship between the amount of suspended load and the corresponding stream discharge (e.g. Leopold, Wolman and Miller, 1964). It has already been shown, however, that the relationship between these two variables is rarely simple for meltwater streams, and data from Bossons also indicate that no bivariate sediment rating curve can be established.

Estimates of the total amount of suspended sediment removed by meltwaters from the Bossons glacier basin have, therefore, been based on the average quantities of suspended sediment measured in April and in July from reach A. Samples

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[1] Rouse's sediment diffusion equation is given by:

$$\frac{Cy}{Ca} = \frac{d-y}{y} \cdot \frac{a}{d-a} \qquad (5.18)$$

where C is the sediment concentration, a is the reference depth above the stream bed, d is the depth of flow, y is any chosen depth, and

$$= V_s/kU_* \qquad (5.19)$$

where V_s is the concentration of a given grain size, k is von Karman's constant, which is determined experimentally, but is usually taken as 0.4 (Simons, 1969; Allen, 1970b) and U_* is the shear velocity (see equation 5.9, page 253).

were selected at hourly intervals during daylight hours over a 10-day period in April, and an eight-hour period in July. A single meltwater sample was considered sufficiently representative of the average stream sediment at the time of sampling because the high degree of turbulence would reduce any vertical gradient in the sediment concentration (see also Pardé, 1933).

(b) Measurements at Bossons and estimates of rates of erosion

Corbel (1959) estimated that 1600 $m^3/km^2/yr$ of suspended sediment was removed from the Bossons glacier basin, compared with 1800 $m^3/km^2/yr$ from Breidamerkurjökull, 2800 $m^3/km^2/yr$ from Hoffellsjökull (see also Thorarinsson, 1939) and 4000 $m^3/km^2/yr$ from Muir glacier, Alaska. Corbel's figures again seem extremely high, mainly because the few samples taken were not representative of the true annual sediment discharge amounts. A much lower estimate of annual sediment removal of 743.37 $m^3/km^2/yr$, equivalent to a layer of 743 mm/1000 yrs, has been obtained from the author's data, using Corbel's formula (see equation 5.14, page 265). This value still seems very high compared with data obtained by Church (1972), who recorded average rates of surface lowering due to removal of sediments in suspension of between 18 and 88 mm/1000 yrs. in the Lewis and Ekalugad basins of Baffin Island. Rates of surface lowering measured by Østrem and Ziegler (1974) were, on the other hand, similar to those at Bossons, ranging from 450 to 1118 mm/1000 yrs. for the Norwegian glacier Erdalsbreen, which covers a similar surface area (11 km^2) to the Bossons glacier.

Equation 5.16 (page 266) introduced earlier by the present author, namely $Q_s = (q_s)(Q_w)/(A)(10^6)$ ($g/km^2/yr$; $m^3/km^2/yr$ = mm/1000 yrs), was considered more accurate than Corbel's formula because it includes estimates of (a) the total meltwater discharge, rather than the depth of a single meltwater stream, and (b) the total discharge of suspended sediment, per unit of meltwater discharge. The proportion of suspended sediment in each

meltwater sample is best expressed in parts per million for conversion to rates of surface lowering in mm.

The average annual amount of sediment removed in suspension from the Bossons glacier has been estimated as 445 $m^3/km^2/yr$, which is equivalent to a surface lowering of 445 mm/1000 years. This value is about a quarter of Corbel's original estimate (1959).

This estimate is based only on the suspended sediment loads in the meltwater streams. Church (1972) and others have shown that the suspended load in proglacial streams may only account for between 4 and 22 per cent of the total sediment load, compared with 50 to 70 per cent measured by Østrem and Ziegler (1974) in Norwegian meltwaters, although the 'mixed load' streams at Bossons (section 5.3), may contain only up to 11 per cent of bedload, and over 89 per cent suspended sediments. The average rates of surface lowering adjusted according to the proportion of the total sediment load accounted for by the suspended sediments range from 4447 mm/1000 years when bedload comprises 90 per cent of the total load, to 494 mm/1000 years, when bedload comprises only 10 per cent of the total load (Table 5.2, page 274).

The removal by meltwaters of sediment at a rate of 494 $m^3/km^2/yr$ from the Bossons glacier basin would represent a total loss of 1.71×10^5 m^3 since 1941. This value can be compared with the total volume of outwash sediments calculated to have accumulated since that date, namely 4.0024×10^4 m^3 (see Chapter 3.4). The total amount of sediment removed from the glacier basin, S_g, is the sum of the amount removed by meltwaters from the valley-train (reach A), S_w, and the amount deposited on the valley-train itself, S_d:

$$S_g = S_w + S_d \qquad (5.20)$$

$$= (1.71 \times 10^5) + (4.0024 \times 10^4)\ m^3/yr$$

$$= 6.40 \times 10^3\ m^3/yr$$

$$= 609.75\ m^3/km^2/yr$$

$$= 610\ mm/1000\ yrs.$$

Table 5.2

Estimates of rates of removal of sediment by meltwaters from the Bossons glacier basin (1971 data)

Type of estimate	Estimated rate of sediment removal ($m^3/km^2/yr$) and rate of surface lowering (mm/1000 yrs) (where $m^3/km^2/yr$ of sediment removal = mm/1000 yrs of surface lowering)
1. Rates of Solution	
(a) Corbel (1959)	
$X = 4E_1T \times 10^{-2}$	200
(b) Smith (1972)	
$X = E_2(T_c + T_m)/10D$	
(i) $T_c + T_m = 10$ (Vivian, 1970; Vivian and Zumstein, 1973)	6
(ii) $T_c=7$, $T_m=5$? (Maizels, 1970)	8
(c) Maizels	
$X = \frac{(q_s)(Q_w)}{(A)(10^6)}$	9
2. Removal of suspended sediment load	
(a) Corbel (1959)	1600
(b) Maizels	
(i) $Q_s = \frac{(q_s)(Q_w)}{(A)(10^6)}$ (q_s in mg/l)	2777 ($kg/km^2/yr$)
(ii) as above, q_s in ppm (cm^3)	445
(iii) Correction of (ii) for bedload = 90% of total	4447
= 10% of total	494
(iv) Correction of (iii) for valley-train deposits	
-bedload = 90% of total	4563
-bedload = 10% of total	610

Table continued...

Table 5.2 (continued)

Type of estimate	Estimated rate of sediment removal etc. (see previous page)	Percentage of total sediment removed
3. Summary of average rates of sediment removal (Bedload = 10 per cent of total load)		
(a) in solution	7.60	1.23
(b) in suspension	444.73	72.05
(c) as bedload (10%)	49.41	8.00
(d) deposited on valley-train	115.51	18.72
Total	617.25	100.00

X = amount of sediment removed annually in solution (mm/1000 yrs)
E_1 = depth of water (dm)
E_2 = total precipitation - evapotranspiration = effective run-off (dm)
T = concentration of sediment (mg/l)
T_c = concentration of $CaCO_3$ in solution (ppm)
T_m = concentration of $MgCO_3$ in solution (ppm)
D = density of rock (g/cm^3)
q_s = total annual discharge of sediment (mg/l, or cm^3)
Q_w = total annual discharge of meltwater (m^3/s)
A = glaciated area (km^2)
Q_s = total amount of sediment removal in suspension (kg, or m^3)

The amount deposited, and preserved, on the valley-train represents about 19 per cent of the total sediment removed from the Bossons glacier (Table 5.2), while over 80 per cent of all sediments eroded from the glacier basin are removed beyond the valley-train by meltwater.

The suspended sediment concentrations of paleo-meltwater streams cannot be precisely determined, although these sediments often form a significant portion of the sediment size distribution in former outwash deposits (see Chapter 6). Nevertheless, the knowledge of their origin, relative proportions, and significance in the rate of surface denudation, will prove useful in later interpretations of the presence of fine interstitial sediments in otherwise coarse, paleo-outwash deposits, such as those in Scotland.

5.5.3.3. Bedload sediments in proglacial streams

The bedload comprises all stream material whose weight is held by the channel bed rather than by the fluid (Bagnold, in Leopold, Wolman and Miller, 1964). Bedload particles move while still maintaining contact as far as possible with the channel floor, through rolling and sliding, and sometimes through saltating. Saltation of the bedload takes on a very significant role in the process of selective transport and sorting of stream sediments when grains of variable density, size and shape are differentially lifted and transported by stream flow. The result of selective sediment transport is often reflected in coarse 'lag' deposits from which the fines have been removed, in small-scale laminations and stratification features, and in larger-scale downstream changes in sediment characteristics (e.g. Jopling, 1963, 1964, 1966, 1969; Brush, 1961, 1965a,b; Harms and Fahnestock, 1965; Happ et al., 1972; and Chapter 7).

The movement of coarse bedload along the channel floor tends to reflect the changes in bed relief and grain size (Church, 1972). The movement of bedload particles is intermittent and transitory, with grains being shifted relatively slowly over short distances along coarse-grained

bars, and more quickly over longer, fine-grained channel segments between successive bars (e.g. Wolman, 1955; Nordin and Beverage, 1965; Keller, 1971; Church, 1972). The total bedload movement is one of the most difficult stream sediment parameters to measure. Few workers have made successful direct measurements of bedload discharge, largely because of problems associated with collecting a representative sample of bedload in motion (e.g. Fahnestock, 1963). The difficulties imposed on direct measurements of bedload transport have resulted in many workers developing empirical or theoretical bedload formulae from which the rates of the total bedload sediment discharge can be deduced (summarized in Graf, 1971). The two most widely adopted formulae are those produced by Einstein in 1942 and 1950, and by Meyer-Peter et al. in 1948. Both formulae should give similar results, and are based on heterogeneous bedloads, i.e. with mixtures of particle sizes (see Appendix 2, and application of formulae in Chapter 7.6).

The movement of bedload material in proglacial streams is characterized by five factors:

(1) The bedload usually represents the most significant portion of the total sediment load in proglacial streams. Church (1972), for example, found that bedload transport accounted for up to 93.6 per cent of the total sediment loads of several proglacial streams in Baffin Island.

(2) Rapid and large variations occur in critical transport flow velocities. The rate of bedload transport can be very high during periods of high flow velocity, but at low regime flow (Froude No. <1.0) little or no bedload transport may occur. Diurnal and seasonal variations in discharge result in diurnal and seasonal patterns of maximum bedload transport movement, in which the coarsest debris is only moved during short-term peak flows each day during the meltseason.

(3) Flood events provide peak flows during which all sizes of bedload material lie below stream competence level. Maximum transport of debris occurs, and the whole

stream bed, including areas submerged by overbank flow, often becomes a mobile sheet of sediment.

(4) Higher rates of sediment transport result from increased turbulence produced by the heterogeneous and poorly sorted nature of the bedload debris which increases the roughness of the channel perimeter.

(5) The abundance of debris covering a wide range of sizes in proximal proglacial outwash deposits means that the meltwater streams rarely lack material of any particular size through its absence in the source material. Debris is available for transport from within the channel (its perimeter and upstream reaches) at all levels of stream competence. The transporting capacity of the stream should therefore more closely approximate the actual, estimated bedload than in stream basins where, for example, the coarser fractions may be absent.

Most of the sediments found in paleo-outwash deposits were, therefore, originally transported along the channel floor as bedload, following the patterns outlined above. The use of the bedload formulae allows the computation of the approximate rates of sediment transport in the former meltwater streams, assuming availability of debris of all sizes, and maximum stream capacity (see Chapter 7). The finer sediments which now form the matrix around the coarse bedload sediments were probably deposited from suspension during a subsequent decline in stream discharge, or were trapped between coarser particles during saltation (see further discussion in Chapter 6).

5.5.4. METHODS OF SEDIMENT DEPOSITION

5.5.4.1. Bar formation and downstream bar growth

Coarse material is selectively deposited on the channel floor during instantaneous falls in tractive force, or is deposited from the collapse of channel banks. The channel becomes increasingly shallow so that relative bed roughness increases. The stream is consequently forced to make compensatory adjustments in other flow or channel

parameters in order to re-establish equilibrium. Additional deposition may therefore take place, possibly accompanied by channel widening to increase the area of the stream cross-section. Experiments by Leopold and Wolman (1957) show that once a coarse gravel lag has formed on the channel floor, bedload transport tends to become concentrated on this raised section of the bed. Moss (1963) describes these accumulations of coarse particles as 'traction clogs'. The bedload, or traction load, is swept up the upstream slope of the lag deposit by turbulent currents and hydraulic lift. The grains then descend down the downstream side, which forms a steeply sloping (over 20°), 'foreset' slope. Particles are frequently graded on this slope, with the coarsest grains rolling, sliding and slipping to the base. Suspended sediment is also added to the foreset slope (e.g. Jopling, 1963) because water shooting across the top of the incipient bar suddenly reaches deeper water in front of the deposit. Flow expands at this point,is diffused and loses its former competence. Particles are selectively deposited out of suspension (e.g. Brush, 1965a), the finest fragments being carried farthest, so that an inclined grading sequence results on the steeply dipping, 'avalanche' foreset faces. The bar deposit is thus enlarged in a downstream direction.

5.5.4.2. Vertical bar growth and overbank deposition

Sediments are built up around the submerged lag deposit to form a mid-channel bar, or point bar on the inside of a channel bend, during increases in discharge. The continued addition of sediment being swept up the back slope of the channel bar causes it to grow vertically towards the water-surface. The top of the bar eventually lies above water-level for most of the time, so that stream flow is diverted round it. Further deposition takes place on the bar top during flood flows, when streams overflow their channel banks. The bar may become stabilized by the gradual colonisation of vegetation (see Chapter 2.6.). Deposition also occurs along the bar edges during floods so that the bar may grow both laterally and

vertically, as well as downstream (e.g. see Williams, 1971). Its growth may be interrupted by erosion if the lateral channels become too narrow for the given discharge. The bar may not increase in size through downstream accretion since the added particles are frequently derived from the upstream sections of the same bar deposit. The bar thus forms a moving depositional system which gradually migrates downstream.

Water flowing over the bar follows an oblique convergent path across the deposit from one bounding channel to the other. Scour by oblique lateral flow dissects many bars to form the characteristic rhomboid shape of transverse bars (e.g. Smith, 1970). Selective deposition occurs as velocity is reduced by greater frictional resistance across the bar top or during the declining stages of overbank, or overbar, flow. Coarse particles are deposited first; fines are later trapped and fall into the interstices to form a matrix (e.g. Happ et al., 1940). The infilling of spaces between the coarser particles with fines is followed by the deposition of a veneer of fines over the coarse gravel surface, described by McDonald and Banerjee (1971) as 'silt-skins' (Figure 5.18, page 281). The fines may also be graded as stream competence declines after widespread sheet-flow across large parts of the outwash surface. Thin fine-grained layers of sediment deposited in this way form 'lineations' in paleo-deposits, often petering out after a short distance as a result of subsequent local winnowing (Figure 5.19, page 281). Flood flows can be the most important depositional agents in areas lying above levels of 'normal' channel flow, for they can transport sediments of most sizes and level out irregularities in surface relief (Jewtuchowicz, 1962a).

5.5.4.3. Some functions of a bar deposit

A bar deposit performs some important functions in the braided fluvial system:

Figure 5.18
'Silt-skin' on mid-channel bar, eroded on downstream edge, Bossons valley-train

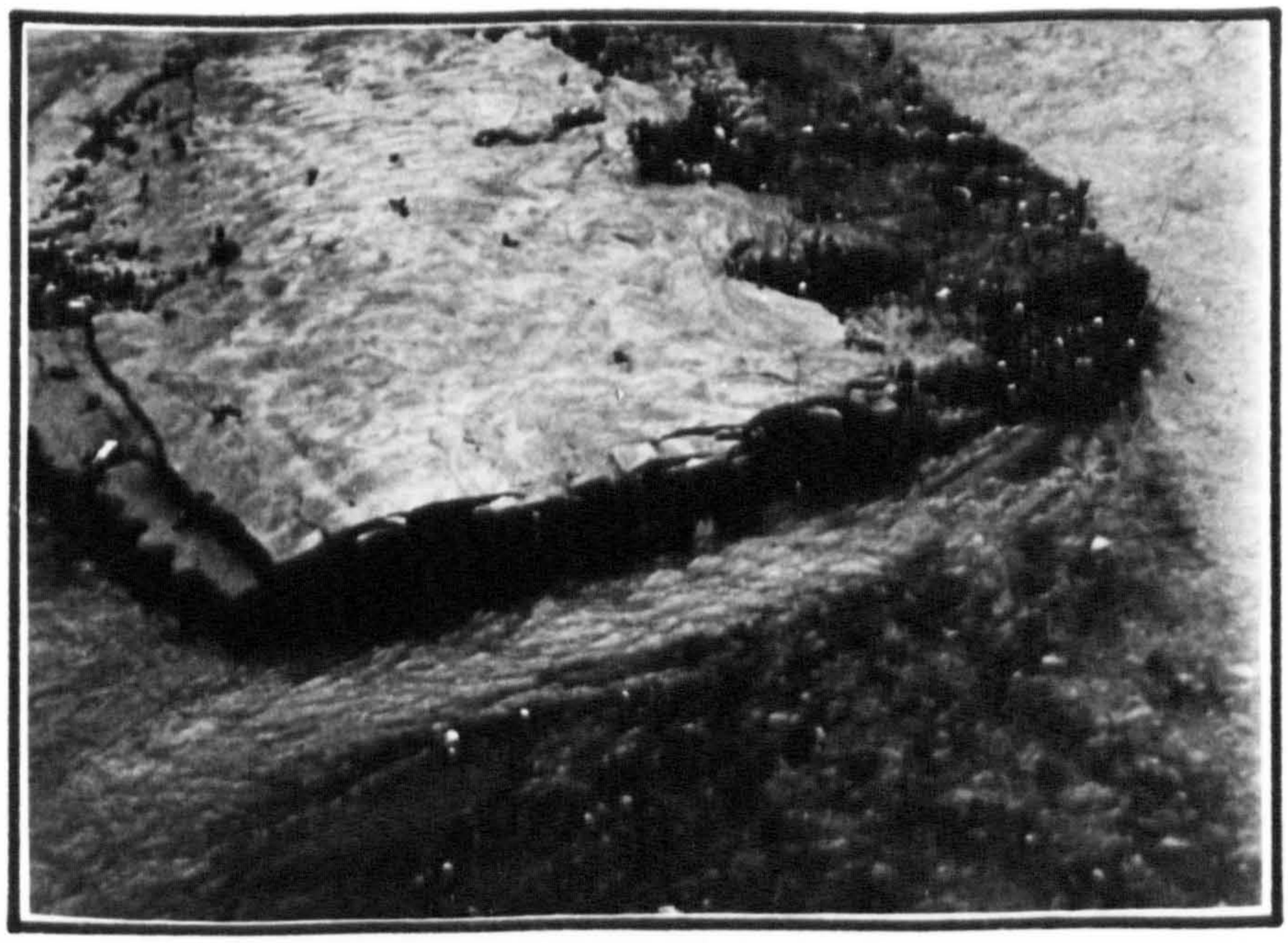
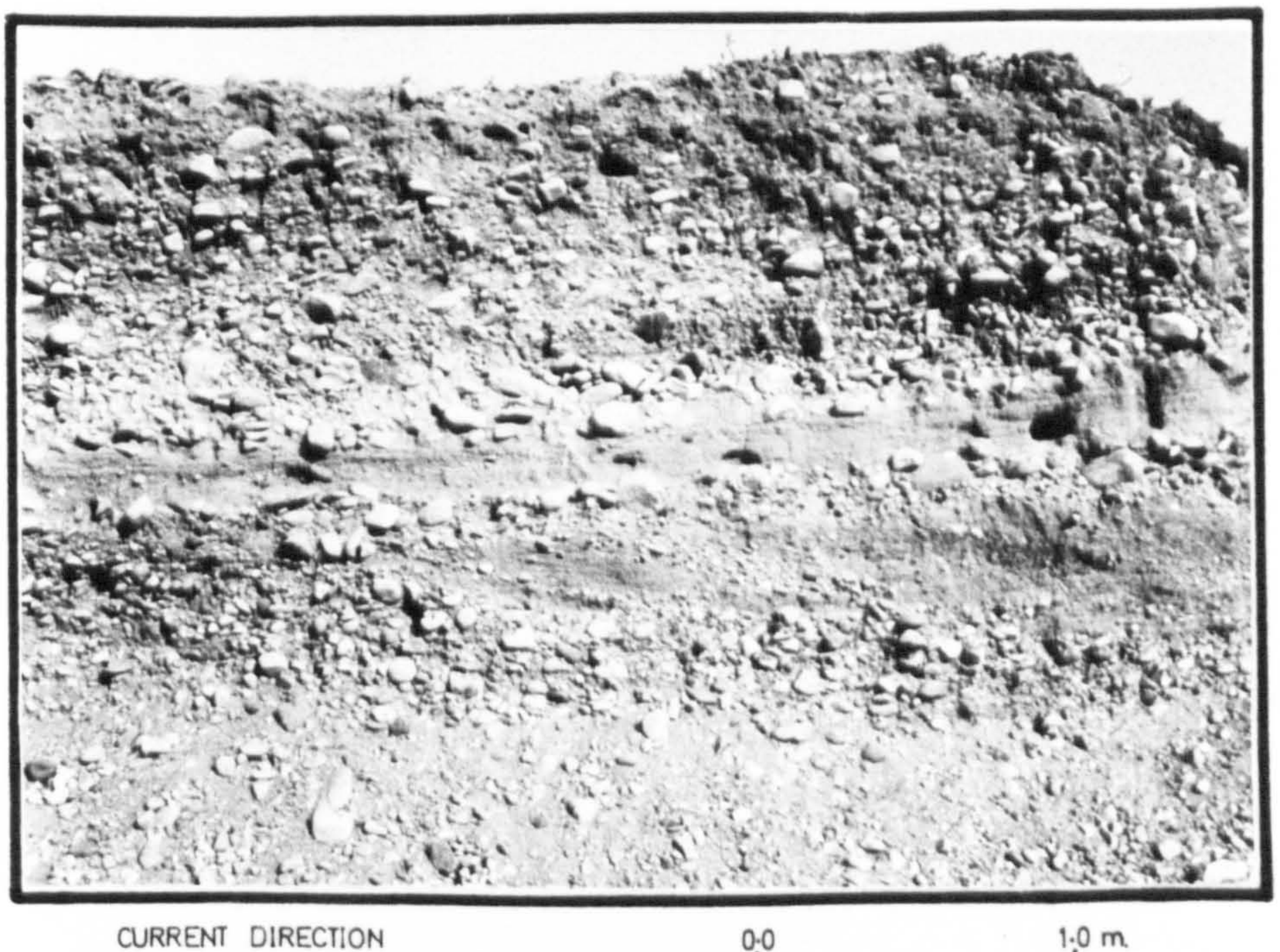


Figure 5.19
Fine sediment lineation in massive, coarse outwash deposits, North Esk gravel pit



(1) It stores extra sediment load until stream competence is again able to transport debris of acceptable dimensions;

(2) It causes the diversion of a single channel into two channels which are more efficient transporting agents, as adjustments in slope are made to cope with sediment load conditions (e.g. Mackin, 1948; Leopold and Wolman, 1957; Dury, 1969); subsequent overbank flooding and bank erosion, lead to rapid lateral shifts of stream courses;

(3) A bar deposit allows the bedload to move downstream by traction along the migrating bar;

(4) Coarse bar-top sediments, originally deposited to form an 'openwork' gravel (containing a high proportion of interstitial spaces) encourage the trapping of fines. The infilling of interstices can reduce bed roughness, so that higher critical velocities are required for erosion to take place (e.g. Sundborg, 1956). The fines nearest the surface may subsequently be removed by selective stream winnowing, but their overall rate of downstream movement will have been considerably delayed (e.g. Meland and Norrman, 1969).

(5) The height of the bar tops provides a local aggradational level which infilling of channels by the vertical accretion of sediments may attain before extensive flood deposition occurs. Bar deposits and channel infill deposits (see below) are not necessarily contemporaneous. Flat bar tops provide an even bedding contact with any subsequent, overlying deposits (see Smith, 1970).

5.5.4.4. Deposition in side channels

The proglacial channel network includes channels which are occupied at successively higher stages, from those containing water almost all the time, to those through which water rarely flows. Side channels usually lie above the level of the main stream and are often only occupied during moderate or high water stages. The water in these side channels often contains higher quantities of suspended sediment than the major streams, largely because

water filters through the channel floor leaving a higher concentration of sediment in the water remaining in the channel (Hjulström, 1952; Schumm and Lichty, 1963).

Water initially enters the side channel during rising flow stage by gradual infiltration, or by water being carried into the side channel by reverse eddies from the main channel. The first water to enter the channel carries clays and fine silts in suspension, but as more water enters the channel and flow depth increases, the sediments become coarser. Through-flow may well be restricted, so that deposition of the coarse sediments results. The channel gradually infills as stream velocities begin to decline and water becomes stagnant (Rust, 1972). Selective deposition occurs as water flows out of the channel either in a downstream direction, or vertically, by sinking through the channel floor deposits. A fining-upward sediment sequence characterizes the infilling of a channel in this way (e.g. Williams and Rust, 1969; Allen, 1970b; Costello and Walker, 1972) and the channel is eventually abandoned (Schumm and Lichty, 1963).

5.5.4.5. Channel infill following stream scour

Channel scour was considered earlier (Section 5.5.2.2.) to represent the initial processes operating during a short-term flood event, during increases in stream discharge. Channel infill represents local deposition during declining stages of a flood event, as the sediment load can no longer be removed as quickly as it is supplied (see Happ et al., 1940). Sediment is usually deposited in the scoured channel floor, excavated only shortly beforehand. The grading of the channel infill deposit may also occur as increasingly finer sediments are allowed to settle as discharge gradually decreases. Subsequent scouring may erode only part of the previous infill, so that the final channel deposit may truncate earlier, similar deposits along erosional contacts, to form a composite infill deposit.

5.5.4.6. Deposition in areas of standing water

A sequence of delta-like deposits, comprising topset, foreset and bottomset sediments, can extend out into small ponds or kettle-holes, eventually filling in the area of deeper water with outwash sediment (see Chapter 3). Pond or deeper water sediment infill appear as pockets of finer sediments intercalated in a coarser gravel sequence in sections of paleo-outwash, often exhibiting some cross-laminar structures (e.g. Jopling, 1966).

5.6. DEPOSITIONAL SEDIMENTARY STRUCTURES

5.6.1. CLASSIFICATION OF BEDFORMS

The type of bedform which develops on the channel floor depends on the nature of the sediment load (particle size, shape, sorting, specific gravity, packing, cohesion), the nature of the fluid (density, viscosity, depth and width of flow, velocity), and the nature of the stream channel (slope, cross-sectional area, alignment, circularity of bends, cohesiveness of banks) (e.g. Schumm, 1960, 1968; Harms and Fahnestock, 1965; Simons, 1969). Flume experiments, carried out by many workers, have shown that fairly close relationships exist between the type of bedform and both sediment characteristics and flow parameters, especially grain size, and depth and velocity of stream flow (e.g. Nordin, 1964; Simons et al., 1965; Harms and Fahnestock, 1965; Jopling, 1966). The grain size of bed particles and the type of bedform each contribute to total resistance to flow, or bed roughness.

Flow régime is a function of the degree of resistance to flow provided by particles and structures on the channel floor at different depths and velocities of flow. Resistance to flow is expressed by a friction coefficient which is a function of mean flow velocity, flow depth and channel slope. Flow régimes are classified into lower, transitional and upper régimes of flow, according to the Froude number (see Appendix 2). The Froude number and

roughness coefficients can be directly related to the type of bedform which develops under particular flow régime conditions (e.g. Nordin, 1964; Jopling, 1966). The lower flow régime is one of 'tranquil' flow (Froude number, Fr $<$1.0), in which silts and fine sands are generally transported by overriding and avalanching over the downstream faces of ripples and dunes. Bed roughness is at a maximum (Manning's 'n' (see Appendix 2) = 0.024 to 0.036; Nordin, 1964; Guy, 1970), and results chiefly from bedform structures on the channel floor. Transition flow occurs during variable conditions and the bedforms which occur may be either ripples or dunes (typically lower flow régime) or plane-bed (typically upper flow régime), according to whichever were the preceding conditions. Upper flow régimes are 'rapid' (Fr $>$ 1.0); turbulence is high, currents and eddies are strong, and the rates of bedload movement are greatest. Particle movement is virtually continuous as grains are moved by hopping, rolling or sliding along the stream bed. This so-called 'high energy environment' is characterised by plane-bed and 'antidune' sedimentary structures (Gilbert, 1914, in Simons et al., 1965). Bed roughness is provided largely by sediment grains, and is therefore at a minimum (Manning's 'n' = 0.012 to 0.018; Nordin, 1964).

The major sedimentary structures which are associated with each flow régime represent the sum of the depositional facies present in a fluvial environment. Recognition of a particular fluvial sedimentary bedform in a paleo-deposit means that other flow parameters, such as the Froude number and Manning's roughness coefficient, can be estimated. The depth and velocity of flow may also be calculated by the use of empirical equations based, in addition, on grain size and channel slope (see Chapter 7 and Appendix 2).

5.6.2. BEDFORMS AND SEDIMENTARY STRUCTURES IN THE OUTWASH DEPOSITS OF BOSSONS AND STRATHMORE

The outwash deposits are composed of a complex series of channel and bar facies, comprising sediments deposited under a wide range of flow régime conditions.

5.6.2.1. Standing water

Standing water facies are only of minor importance at Bossons and are present locally in temporarily abandoned channels, in sheltered parts of the outwash surface, and on the floors of kettle-holes. Standing water, with only occasional periods of through-flow, provides a very 'low energy' environment in which silts and clays fall out of suspension to form a laminated bottom deposit, sometimes containing 'dropstones' which have melted out from ice rafts on the water surface (e.g. Ovenshine, 1970). The fineness of the particles increases the capacity of the whole deposit to absorb water so that it can form a structureless, mobile, "quick" deposit, in which the thixotropic silty clays are veneered with a layer of 'plastic mud', 'Silt volcanoes' often erupt through the waterlogged deposit, forced up by hydrostatic pressure as the underlying layers of clay are being compacted (Figure 5.20, page 287).

This type of sediment is rarely preserved in section unless it had been well consolidated before further, overlying sedimentation took place (e.g. Williams, 1971). Some of the finely laminated silts in the North Esk outwash deposits were sufficiently saturated to have become mobile and now exhibit relict flow structures (Figure 5.21, page 288). Subsequent consolidation and burial by coarse fluvioglacial deposits allowed these paleo-sediments deposited in standing water, to be preserved in part.

5.6.2.2. Ripples, dunes and cross-bedded structures

Low stream velocities, relatively shallow depths and fine particle sizes characterize the low flow regime in which ripples and dunes develop on the stream bed (e.g. Menard, 1950). Ripples and dunes are frequently concentrated along the channel floor in distal portions of the Bossons valley-train (Figure 5.22, page 289). Bar tops are often topped by 'silt-skins' and ripples which formed during the declining stages of flood flow over the bar.

Figure 5.20
Deposition in a shallow pool,
Bossons valley-train

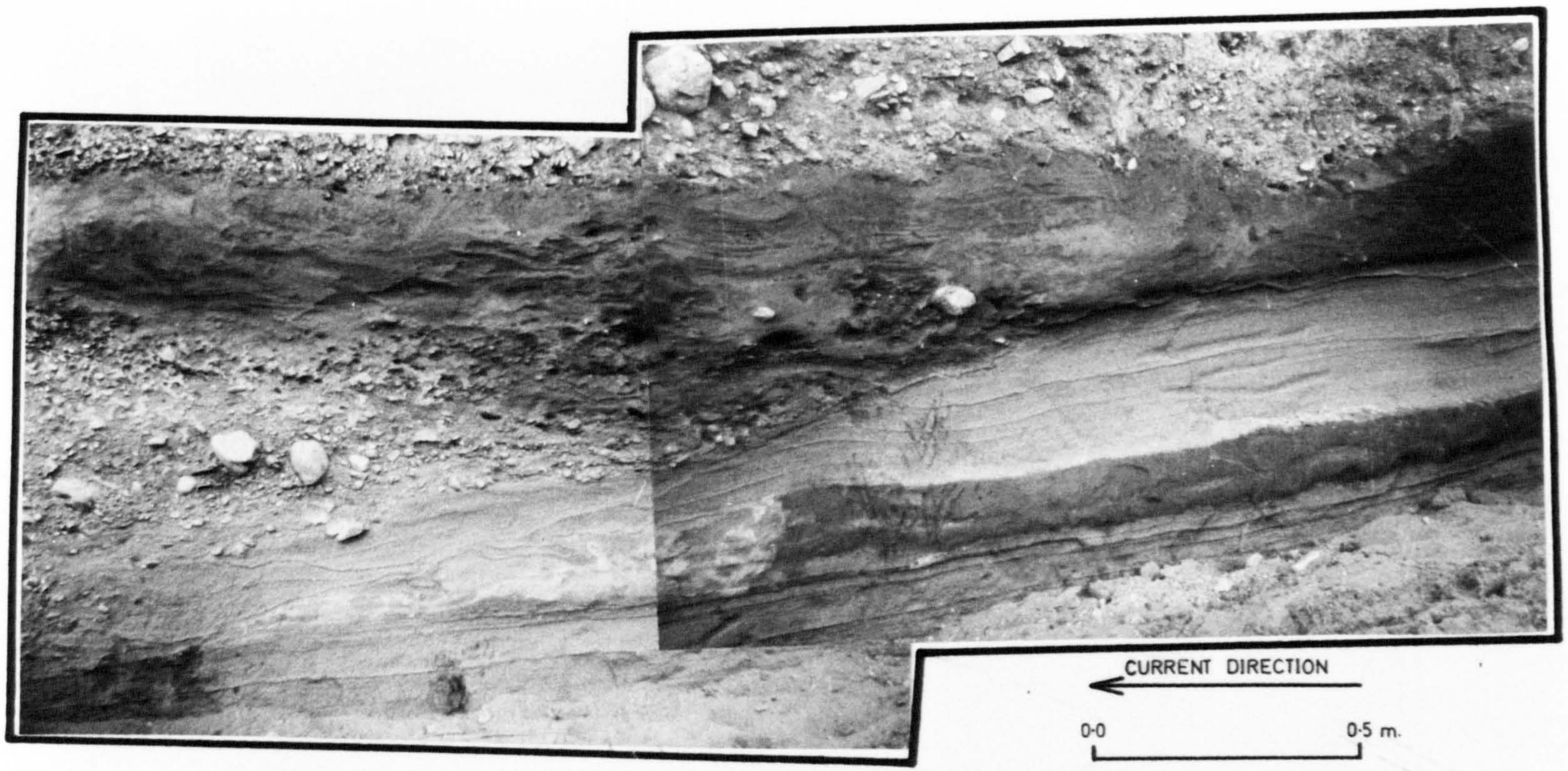


Figure 5.21 Relict flow structures in formerly saturated, fine-grained, ice-marginal sediments, North Esk kame system near Meikle Tullo

Figure 5.22
Ripples on the bed of a meltwater stream,
Bossons valley-train

Ripples and dunes have been partially preserved in the paleo-outwash deposits and characteristically exhibit the cross-laminated or cross-bedded structures which developed during their formation (e.g. Figure 5.23, page 291 ; Figures 5.24 and 5.25, page 292). The axis of each ripple or dune lies perpendicular to the local current direction, and each bedform initially contained a bottomset, a foreset and a topset series of laminations produced by the downstream transport of grains up the incline of a bed irregularity or clump of grains (Menard, 1950), and the overshooting or avalanching of grains across the downstream face. Ripples and dunes are streamlined bed configurations which provide a means of sediment transport along the bed, even at low flow velocities. The topset is sometimes removed as the bedform migrates or as erosion takes place when discharge increases. The steeper foreset laminations advance to overlap the bottomset, to produce a series of cross-laminations. Cross-laminated or cross-bedded structures can be progressively built up during stream aggradation (Figure 5.23, page 291), resulting in a complex series of partially truncated laminar structures.

A considerable amount of work has been published on the types of internal structures of ripples and dunes (e.g. Illies, 1949; Pettijohn, 1962; Simons et al., 1965; Harms and Fahnestock, 1965; Jopling and Walker, 1968; Allen, 1970a; and many others). Ripples and dunes have been differentiated according to their constituent grain sizes, their dimensions on the channel floor, and the velocity and depth of stream flow during formation. Dunes are coarser-grained (>0.6 mm), larger in dimension (>4.5 cm) and are formed at higher velocities than ripples (e.g. Nordin, 1964; Simons et al., 1965). The major types of cross-bedded structures have also been differentiated according to their degree of curvature, symmetry, variability of height and spacing, cross-sectional profiles and grain sizes (see Harms, 1969). Several major groups have been observed in localized areas of the Bossons outwash

Figure 5.23
Cross-bedded sands in paleo-outwash deposits,
South Esk gravel pit

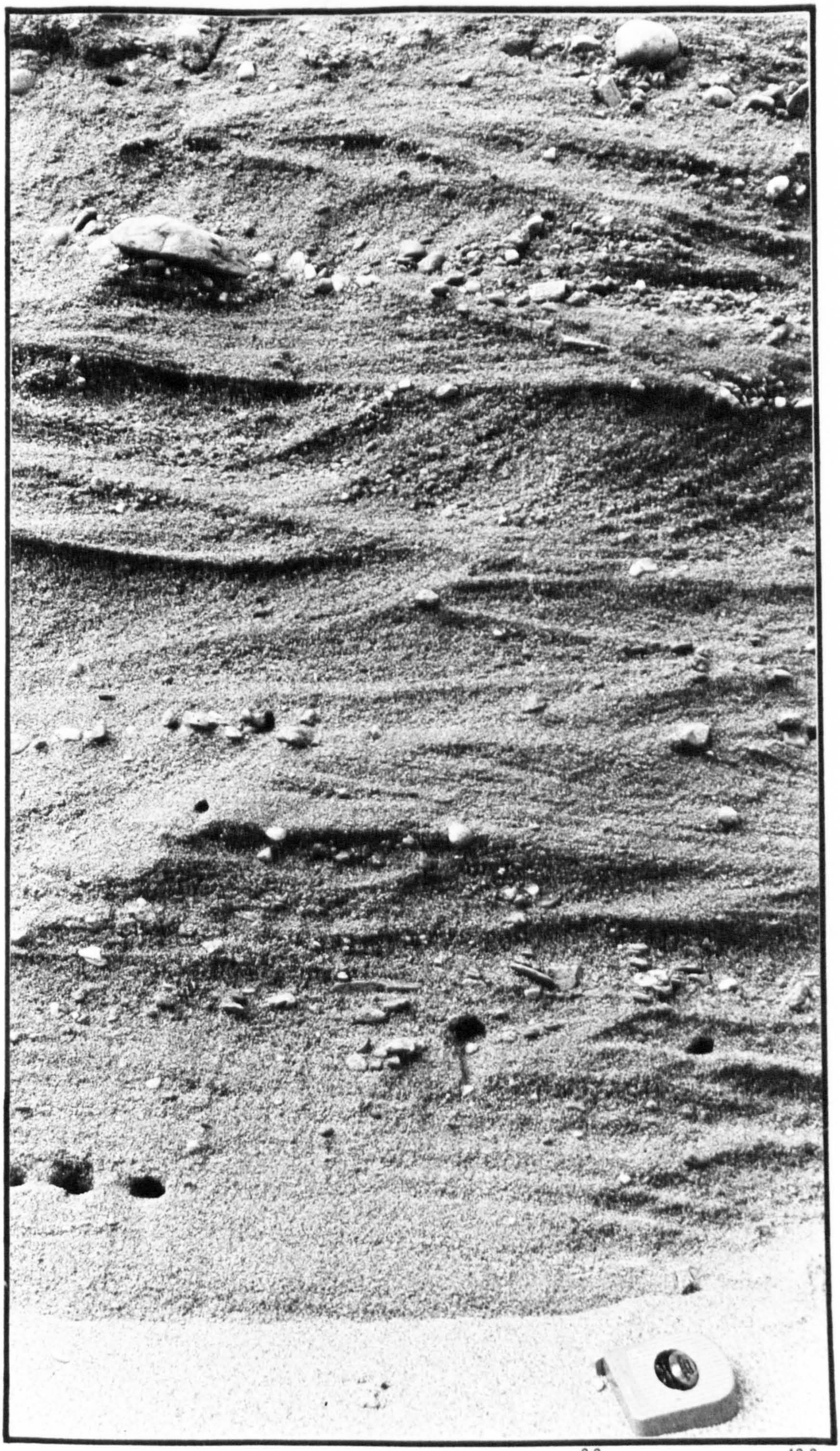


Figure 5.24
Planar cross-bedding, overlain by plane-bedding, in coarse paleo-outwash sands, South Esk gravel pit

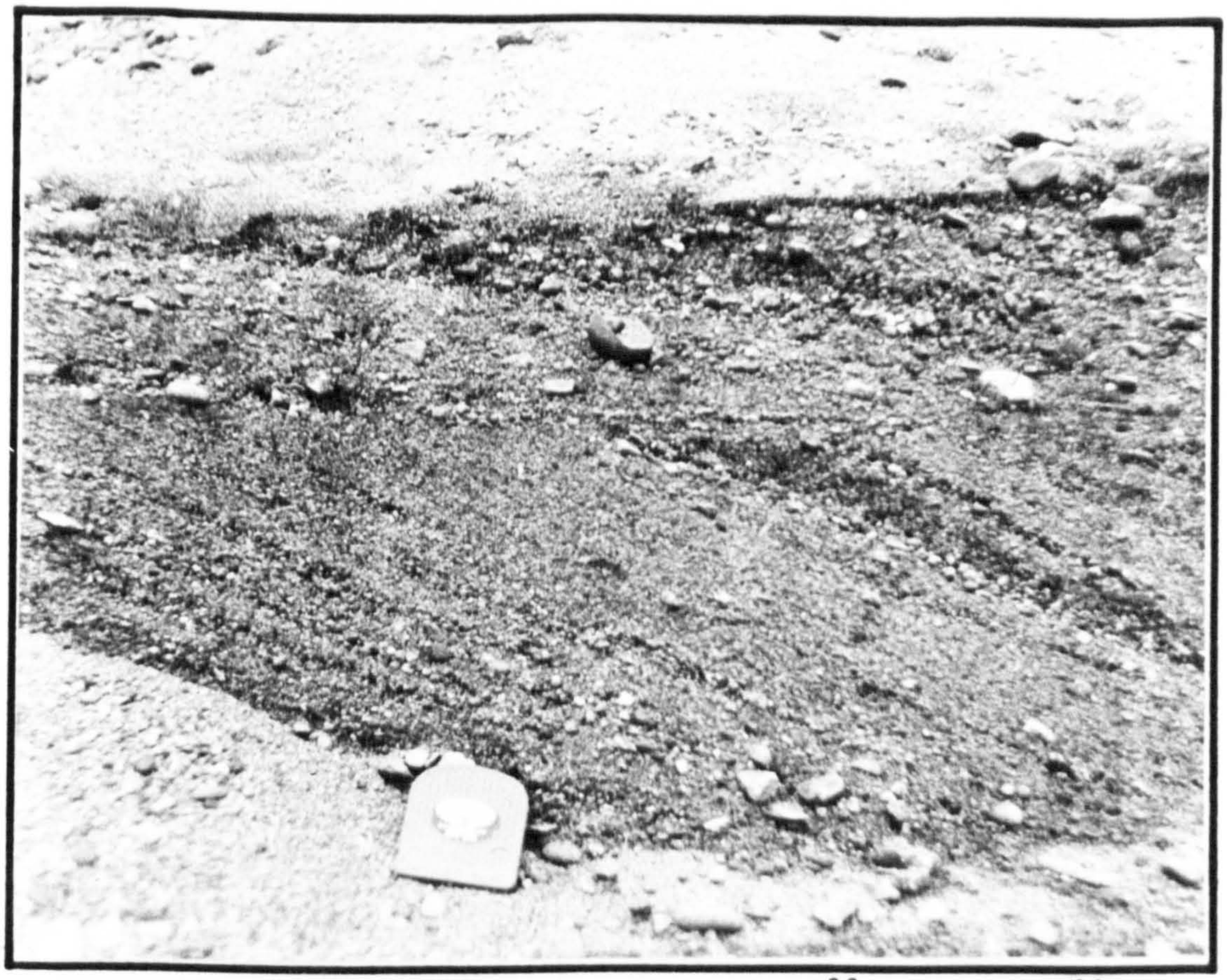
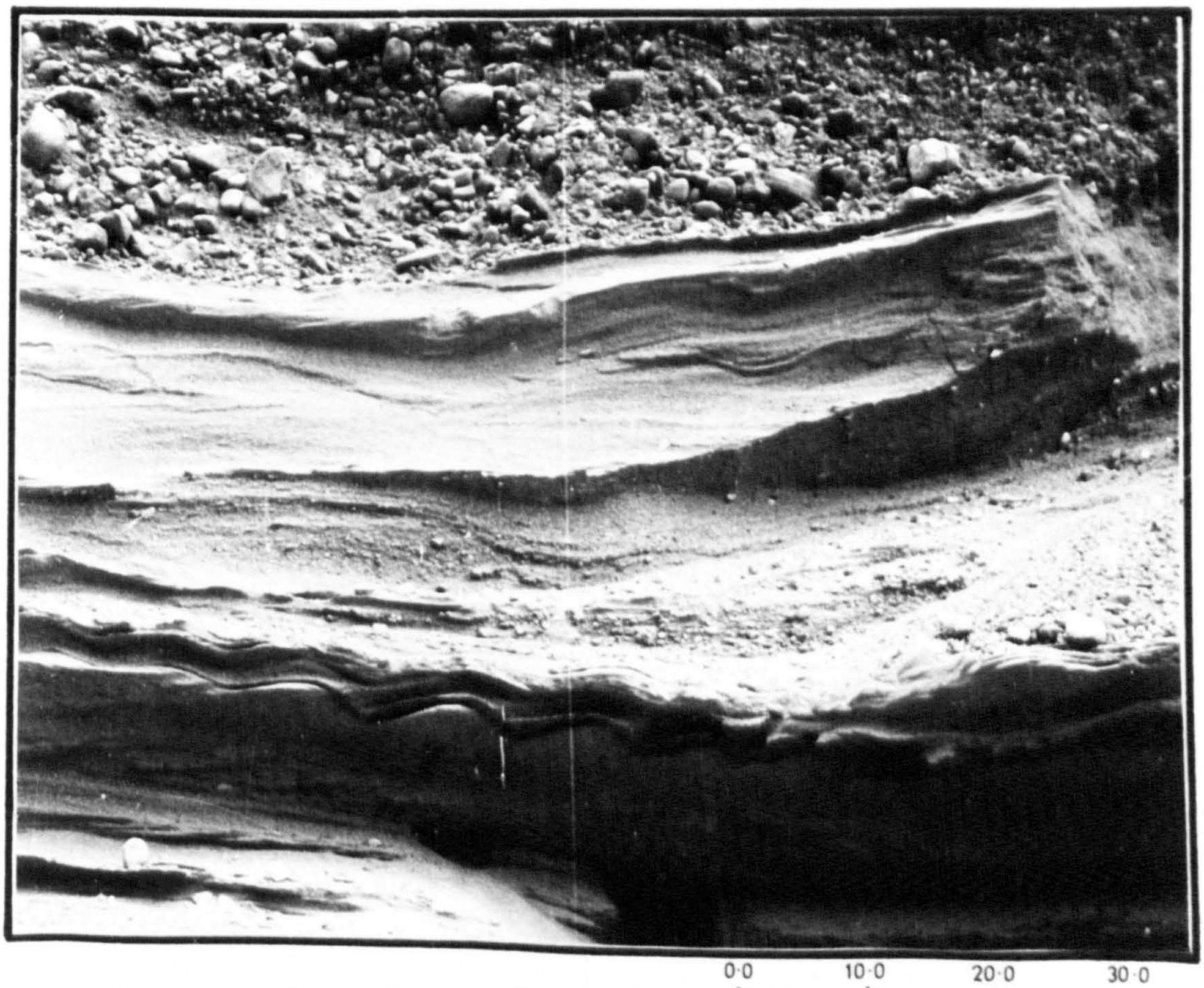


Figure 5.25
Ripples overlain by an erosion surface in paleo-outwash deposits, sand pit, Edzell golf course



deposits and in isolated sections of the paleo-deposits.

The most extensive cross-bedded structures in the paleo-deposits are straight and parallel, and therefore classified as planar or torrential cross-stratification (Figure 5.24). Planar cross-stratification is formed by the migration of larger ripples, 'sand waves' or dunes and is usually composed of gravelly sand or sandy gravel associated with the movement of transverse bars (Smith, 1970). The direction of the maximum dip of the laminations provides a useful indicator of current flow direction (Harms, 1969). Over twenty clear lenses and lineations of these cross-bedded sands and gravels have been preserved in a 300m face exposed in the North Esk gravel pit (Figure 5.26, page 294). The angle and direction of maximum dip has been determined from twenty of the cross-bedded structures, although they are frequently less than 10cm thick (Figure 5.27a and Table 5.3; compare Roep et al., 1975). Dips of between $8^{o}30'$ and $37^{o}15'$ were recorded (Table 5.3, page 297 , and Figure 5.27b, page 295). The former current directions depositing cross-bedded structures in the North Esk gravel pit ranged from N.75^{o}E. to N.150^{o}E., with a modal current direction of N.130^{o}E. to N.150^{o}E (Figure 5.27c; $\pm 5^{o}$, Harms, 1969), a direction which is almost parallel to the pit face and to the present River North Esk.

Ripple and dune cross-laminations which are slightly curved and either symmetrical (Jopling and Walker's (1968) 'ripple-drift cross-lamination', Type B) or asymmetrical ('sinusoidal ripple lamination') were present in the North Esk outwash deposits (Figures 5.25 and 5.28). The site (Edzell Golf Course, site E47) lies towards the north of the probable location of the former Westwater-North Esk confluence zone, and at a slightly higher elevation. These laminations were formed at lower velocities and of finer sediments than farther south or east in the major outwash tracts.

Trough or festoon cross-laminations, which represent the cross-sections of ripple and dune structures, were found extensively in the downstream exposures in the South Esk

FIGURE 5.26 THE CHIEF SEDIMENTARY STRUCTURES OF THE NORTH ESK GRAVEL PIT

Based on a section exposed in August 1974

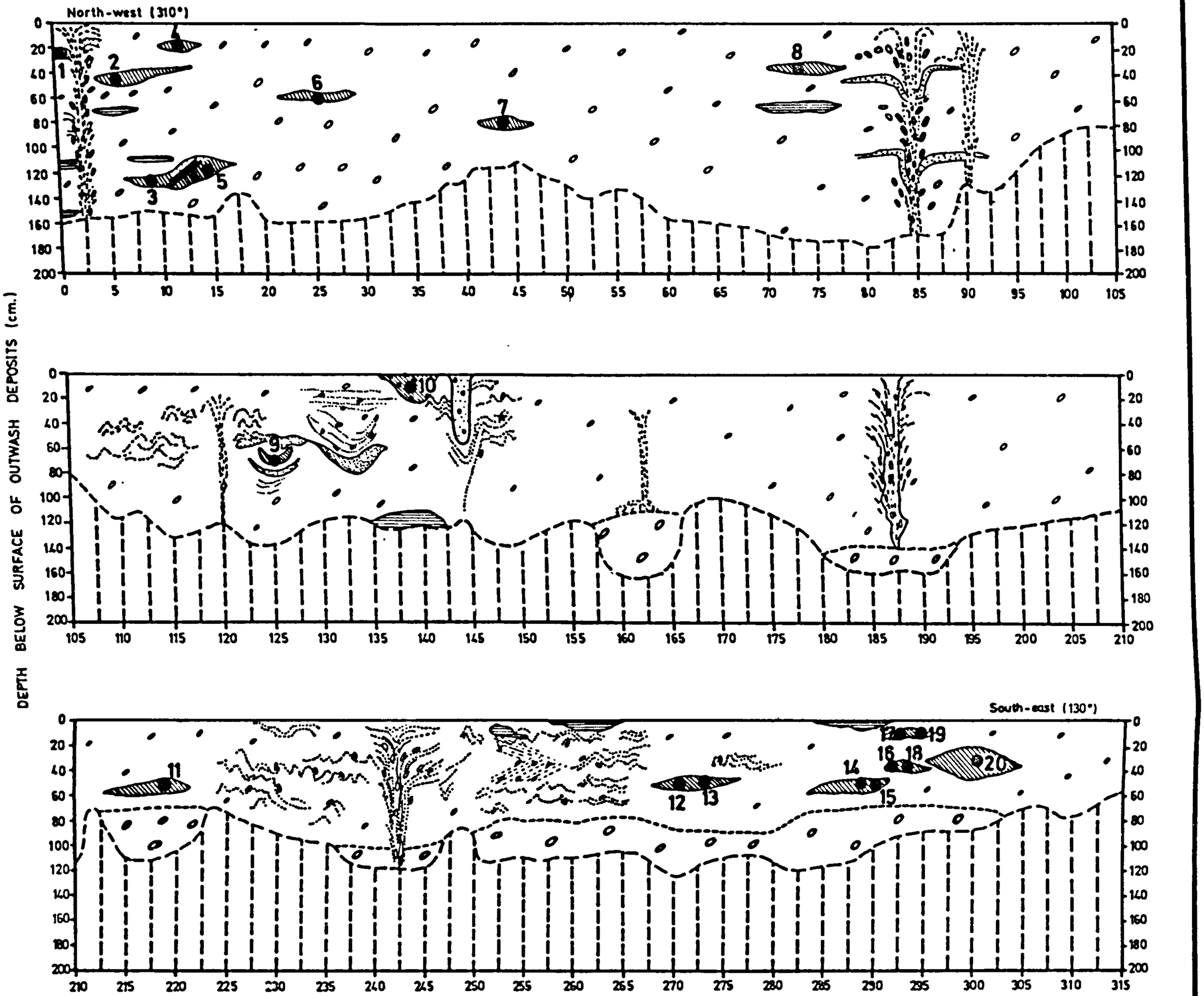
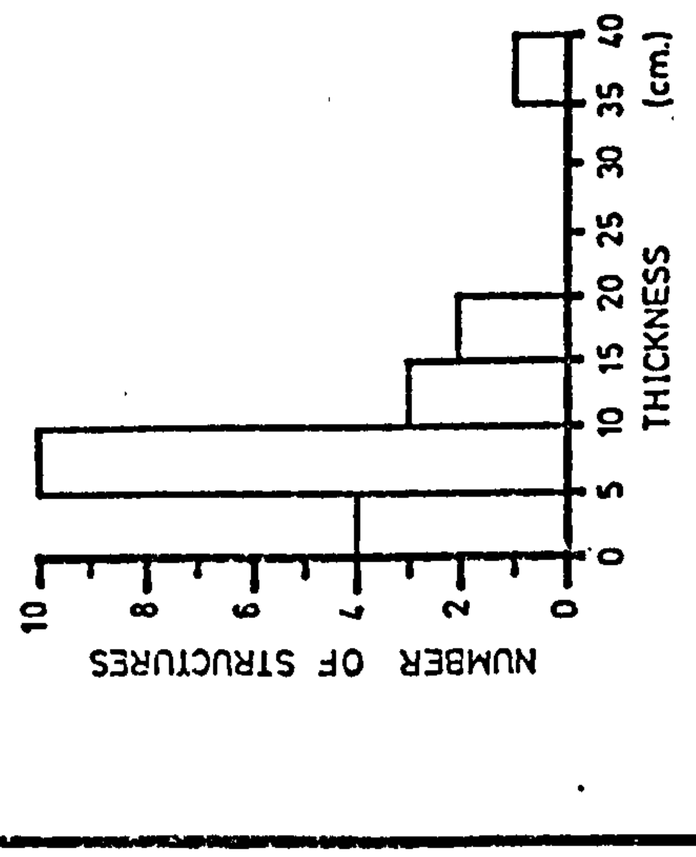


FIGURE 5.27 THICKNESS AND DIP OF CROSS-BEDDED STRUCTURES IN THE NORTH ESK GRAVEL PIT

(a)Thickness of cross-bedded structures (b)Angle of dip of cross-bedded structures (c)Direction of dip of cross-bedded structures



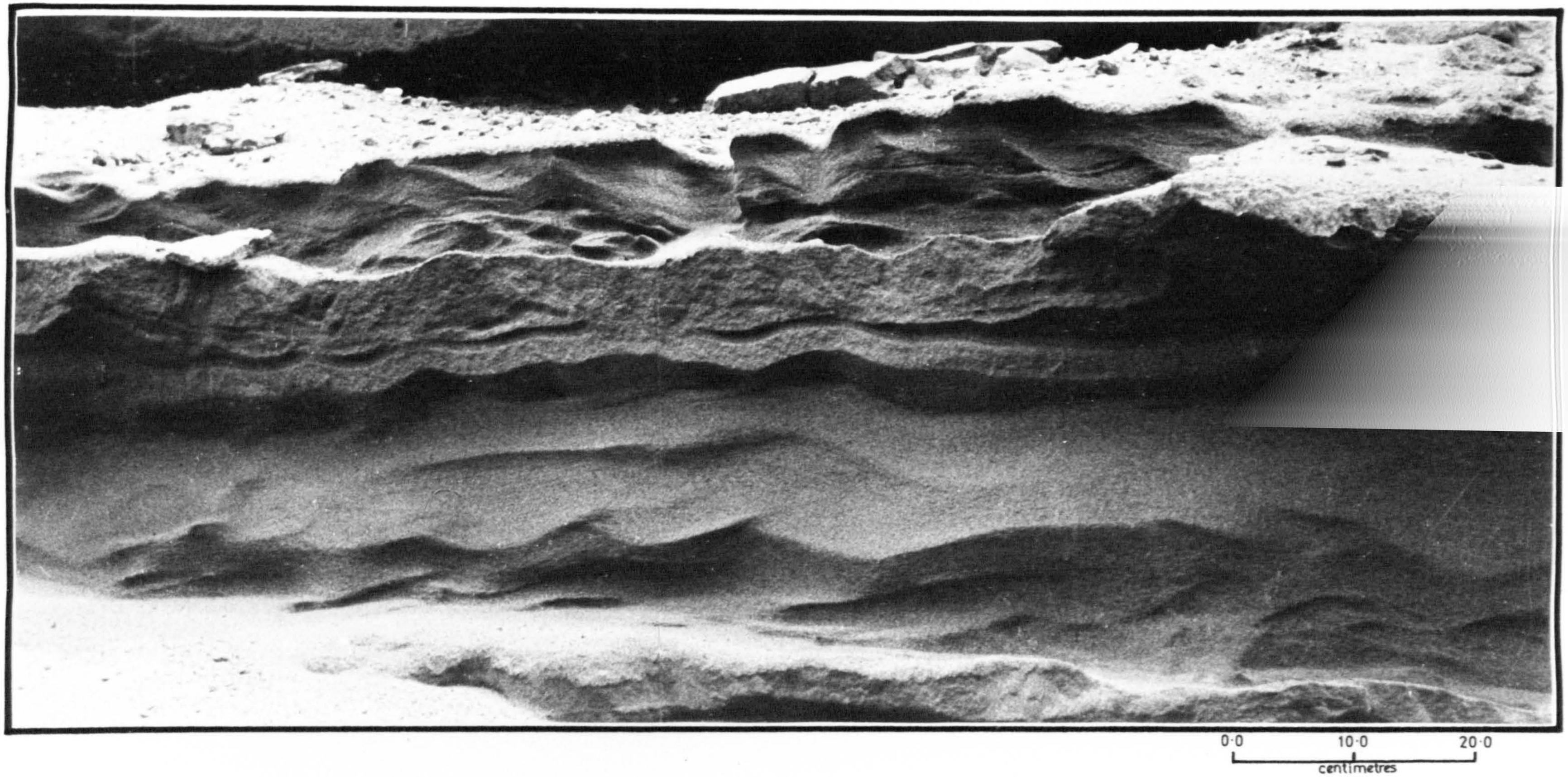


Figure 5.28 Ripple structures in paleo-outwash deposits, sand pit, Edzell golf course

Table 5.3

Cross-bedded structures in the North Esk gravel pit

Cross-bedded structure	Composition s=sand g=gravel	Thickness (cm.)	Angle of dip	Direction of dip (North=0°)	
1	s	5	29° 05'	126°	
2	s	4	34° 40'	145°	
3	s+g	10	27° 50'	127°	
4	s	9	27° 50'	125°	
5	g	7	37° 15'	128°	
6	s	8	23° 55'	136°	
7	s	18	16° 40'	133°	
8	s	9	13° 45'	132°	
9	s	16	23° 25'	153°	
10	s+g	36	16° 10'	135°	
11	g	9	29° 50'	140°	
12	s+g	13	18° 00'	95°	
13	s	13	34° 30'	140°	
14	s	8	23° 50'	75°	
15	s	2	33° 20'	145°	
16	s	5	12° 00'	325°	(antidune?)
17	s	6	25° 20'	135°	
18	s	3	8° 30'	140°	
19	s	6	32° 15'	125°	
20	s	3	-	100°	

300m downstream ⇓

outwash deposits (Figure 5.29, page 298). The large-scale trough cross-laminations accumulated in scour hollows and channels which were subsequently infilled during declining flow stages (e.g. Williams, 1968). The smaller-scale trough cross-laminations were probably deposited in scours formed by eddies downstream of migrating ripples. Continued partial erosion and subsequent channel infill, especially during progressive downstream and/or lateral migration, has led to the development of a thick, composite series of trough cross-laminations, lying perpendicular to the channel axis (Figure 5.30, page 299).

Cross-bedded structures may also be found on the margins of coarse bar cores, in abandoned channels traversing the bar top (e.g. Sundborg, 1956; Eynon and Walker, 1974), and also as delta-like series of deposits

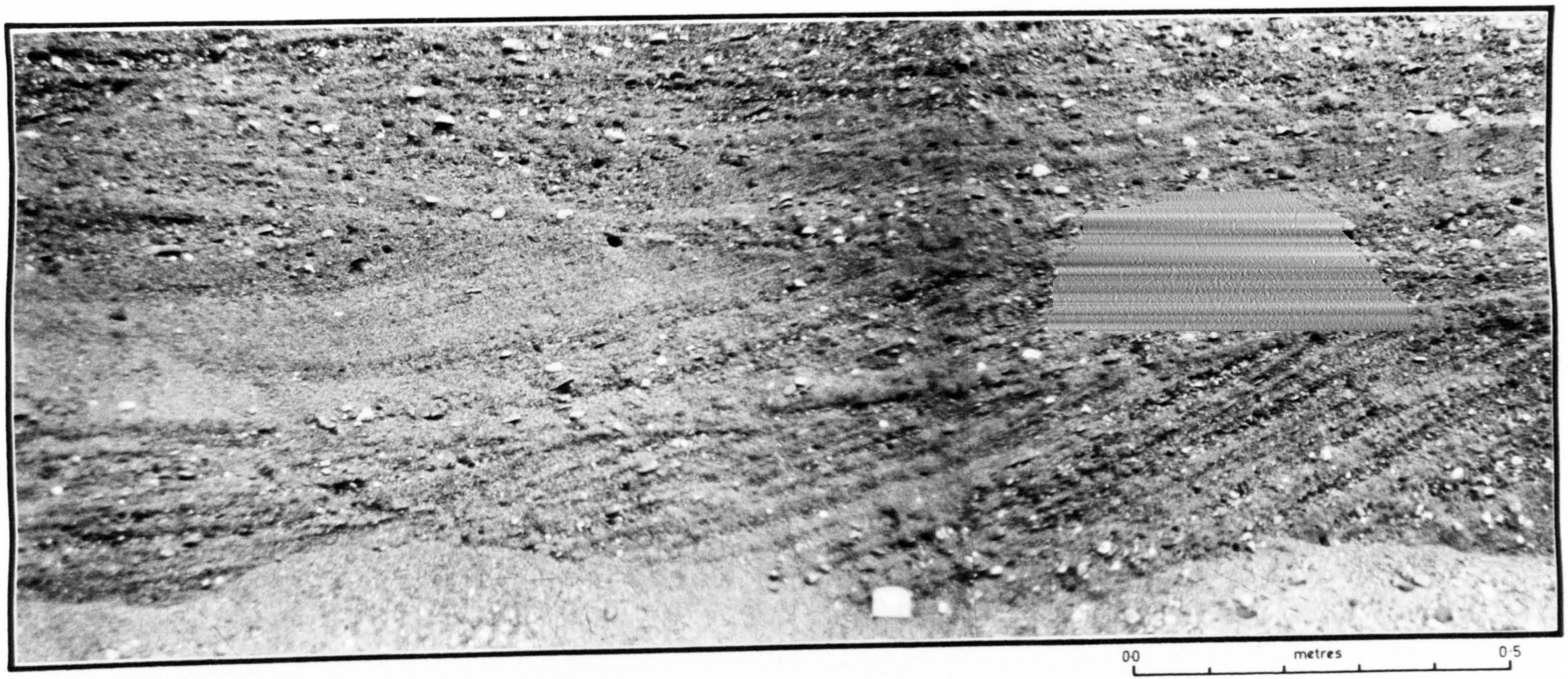


Figure 5.29 Trough cross-bedding (left) merging laterally with planar cross-bedding (right) of channel margins in paleo-outwash sands, South Esk gravel pit

Figure 5.30
Trough cross-bedding, South Esk gravel pit

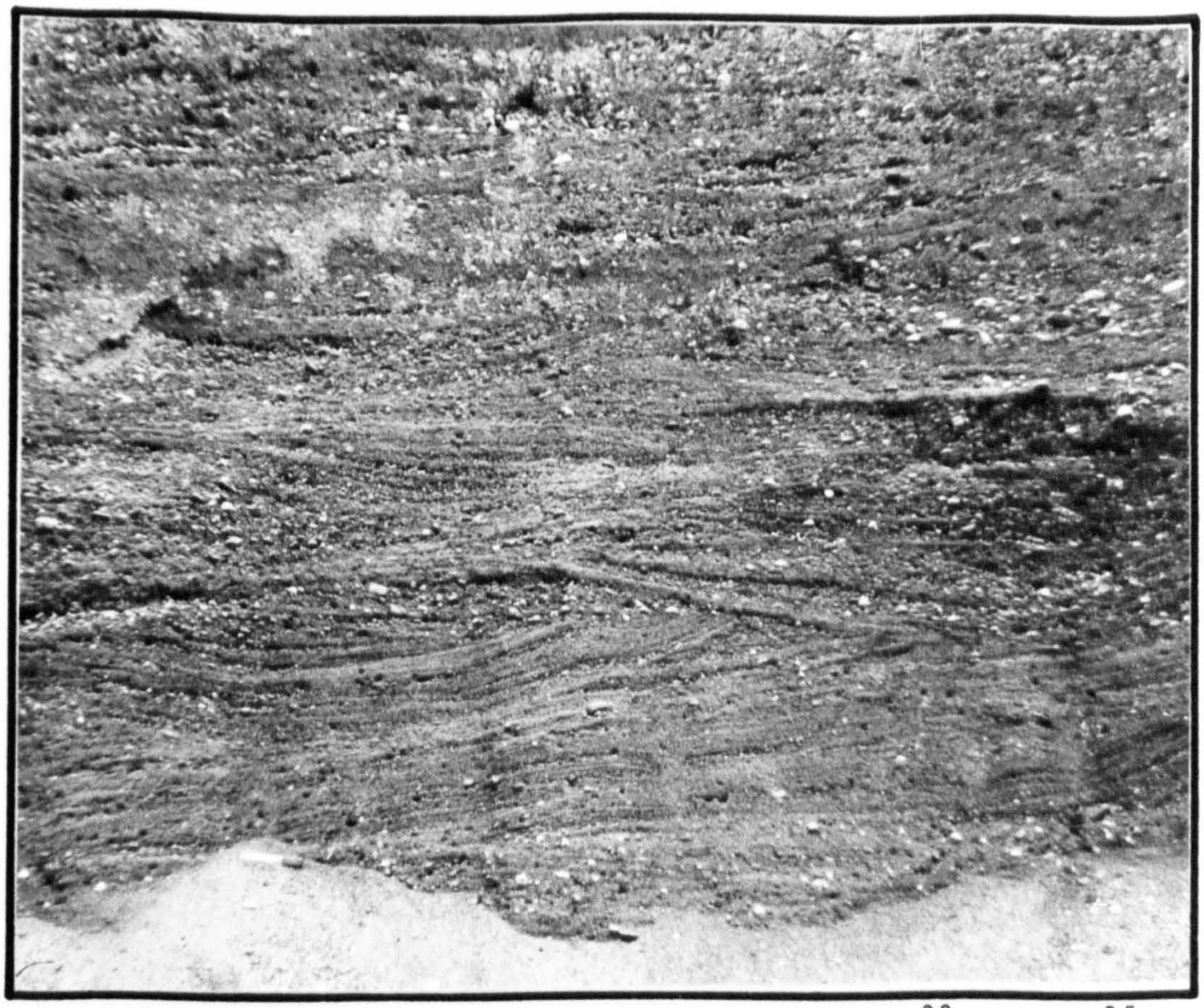
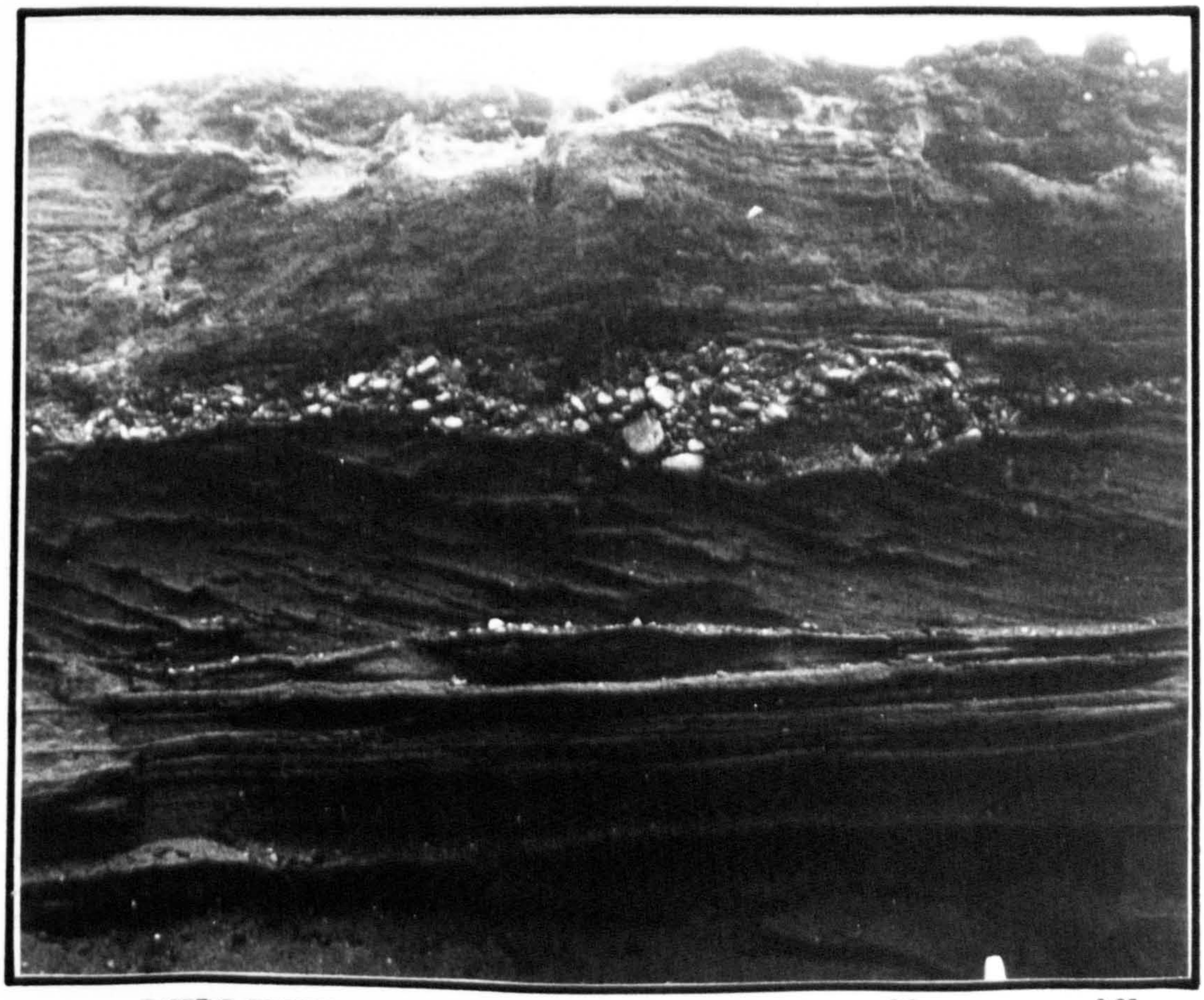


Figure 5.32
Coarse channel infill truncating cross-laminated and plane-bed sands, South Esk gravel pit



which accumulate in areas of standing water such as the kettle-holes (see Figure 5.31, page 301).

Many of the cross-bedded structures in the paleo-outwash deposits are overlain by coarse sediments deposited at high flow velocities after a period of channel scour. The upper boundaries are usually conspicuously erosional (e.g. Figure 5.25, page 292 , and Figure 5.32, page 299).

5.6.2.3. Plane-bed sedimentation

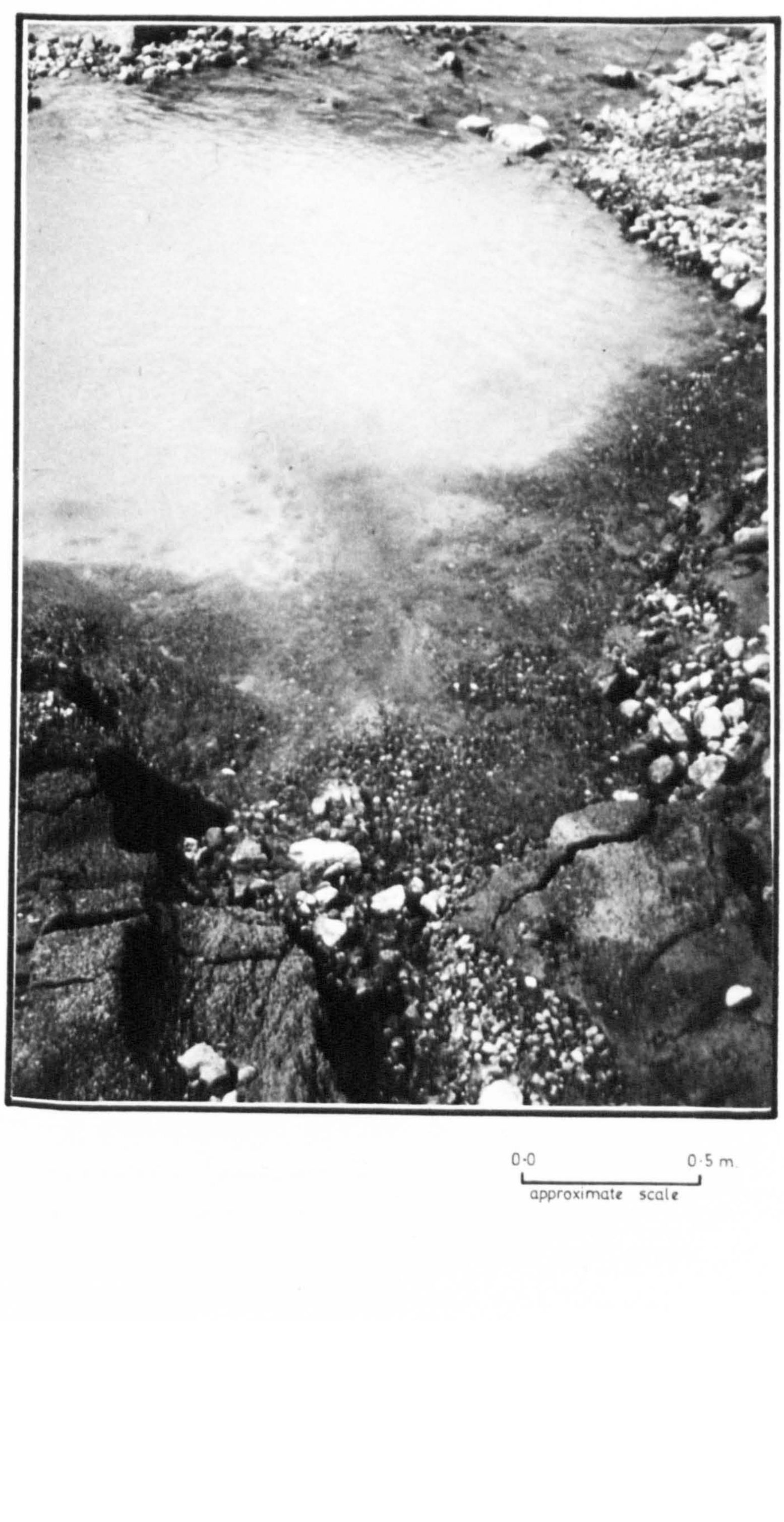
(1) Horizontal bedding of sands

Horizontally bedded, sandy strata have been widely observed in these and other Pleistocene outwash deposits (e.g. Figure 5.24, page 292 ; Figure 5.32, page 299 ; Figure 5.33, page 303). The sands were deposited during a period of almost continuous 'sheet-flow' of the stream bed in upper flow régime conditions. The sands are often poorly sorted and crudely stratified as a result of rapid deposition during high rates of bedload transport. The strata frequently lie on a sharp horizontal basal plane with few bed obstructions, representing a former bar surface (e.g. Williams, 1971). Coarse, plane-bedded sands are most frequently associated with longitudinal bar deposits (Smith, 1970). The upper surface may be a distinctly erosional contact, truncated by trough and channel infills.

(2) Massive deposition of gravels and imbrication of coarse particles

Crudely horizontal or massive, structureless coarse gravels are widespread in all the outwash deposits investigated, and have also been reported from other proglacial deposits by many workers (e.g. Seret, 1965; Williams and Rust, 1969; Church, 1970, 1972; McDonald and Banerjee, 1971; Eynon and Walker, 1974). Their origin is attributed to 'high energy' sheet-flow of gravels, with little bed resistance across the outwash surface during flood discharges (Manning's n = 0.012 to 0.015, Simons, 1969; and see Appendix 2). The poorest bedding results from the most rapid rates of deposition and when few fine fractions are present (e.g. Schumm, 1960a,b; Williams and

Figure 5.31
Deltaic sedimentation in a kettle-hole,
the Bossons valley-train



Rust, 1969). Most of these gravels form bars, the cores of which are coarsest, while the top sediments are reworked during floods. Fluctuations in stream competence cause local deposition of finer sediments producing numerous fine-grained interfingering ribbons, lenses and discontinuous horizons.

Rapid movement of coarse bedload has resulted in the clasts possessing a preferred orientation and dip which is often, therefore, the only distinct internal structural pattern present in the coarse outwash deposits (e.g. Figure 5.35, page 304). The flatter, slab-shaped clasts are rolled over the stream bed until their long axes lie parallel to the current and an upstream dip is attained (Twenhofel, 1926; Lane and Carlson, 1954; Laming, 1966). Hydraulic lift is hindered by clasts lying in this position, as the drag forces acting downwards on the bed particles effectively hold them down on the bed (Laming, 1966; Allen, 1970a). Fine sediments, or further coarse material, may rapidly infill the interstices and the original particle dip is maintained in the aggrading deposit. The result is either 'isolate imbrication' when there is a low ratio of clasts to matrix (Laming, 1966; and Figure 5.34, page 303), or 'contact imbrication' when numerous clasts are 'shingled' with an upstream dip of 15 to 30° (Laming, 1966; Blatt et al., 1972) (Figure 5.35, page 304).

Many other workers have observed an imbricated particle structure in coarse fluvial and fluvioglacial deposits (e.g. Twenhofel, 1926; Cailleux, 1938; Wadell, 1936; Krumbein and Pettijohn, 1937; Krumbein, 1939; Schlee, 1957a; Doeglas, 1962; Fahnestock, 1963; Moss, 1963; Laming, 1966; Church, 1970; Boothroyd, 1972). Measurements of imbrication should be based on three-dimensional exposures (but see Schlee, 1957; Laming, 1966), and the resultant direction of imbrication can then be used as an additional indicator of the direction of paleo-flow currents (see also Costello and Walker, 1972).

Figure 5.33
Plane-bed gravels overlain by coarse paleo-outwash deposits, North Esk gravel pit

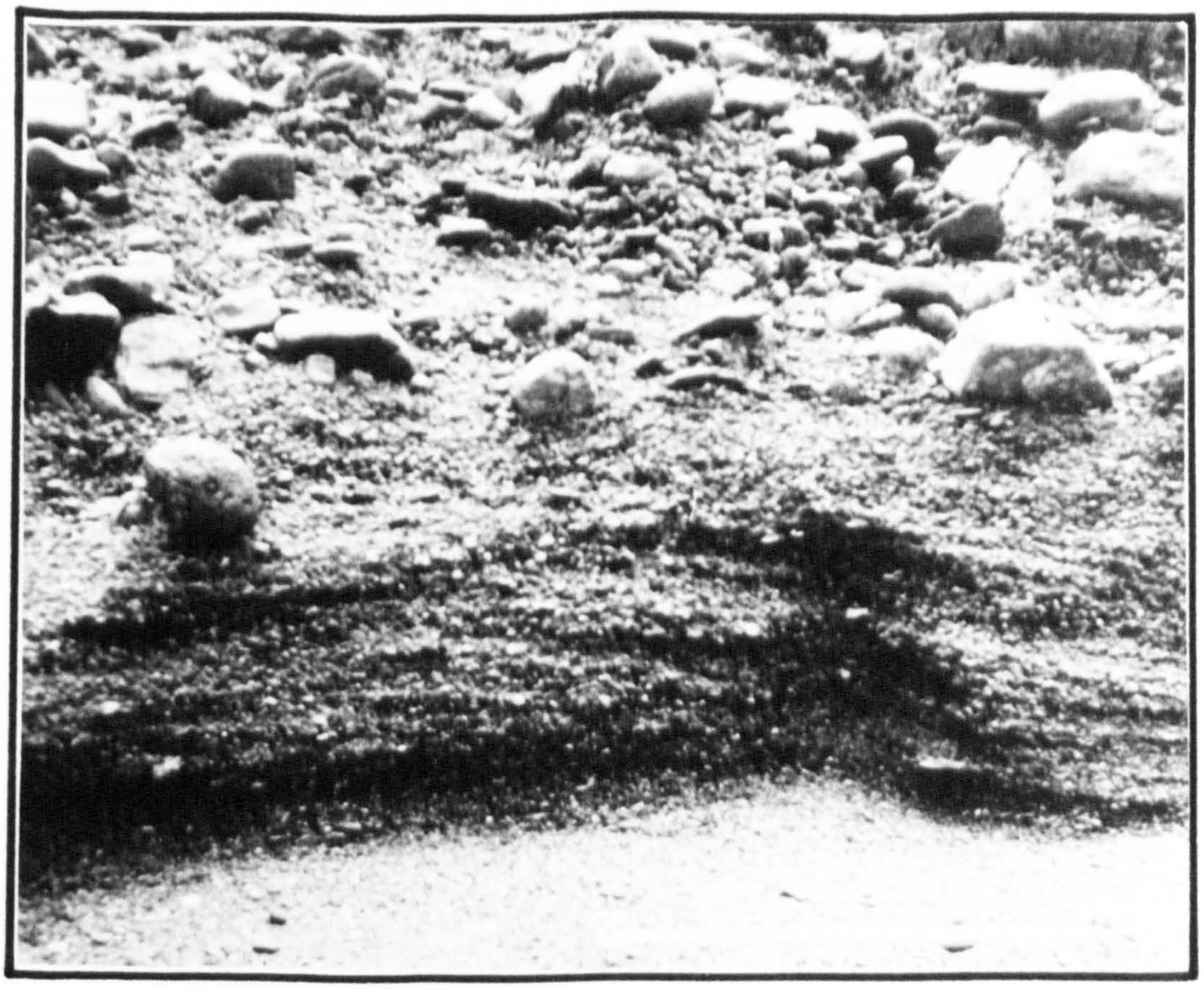
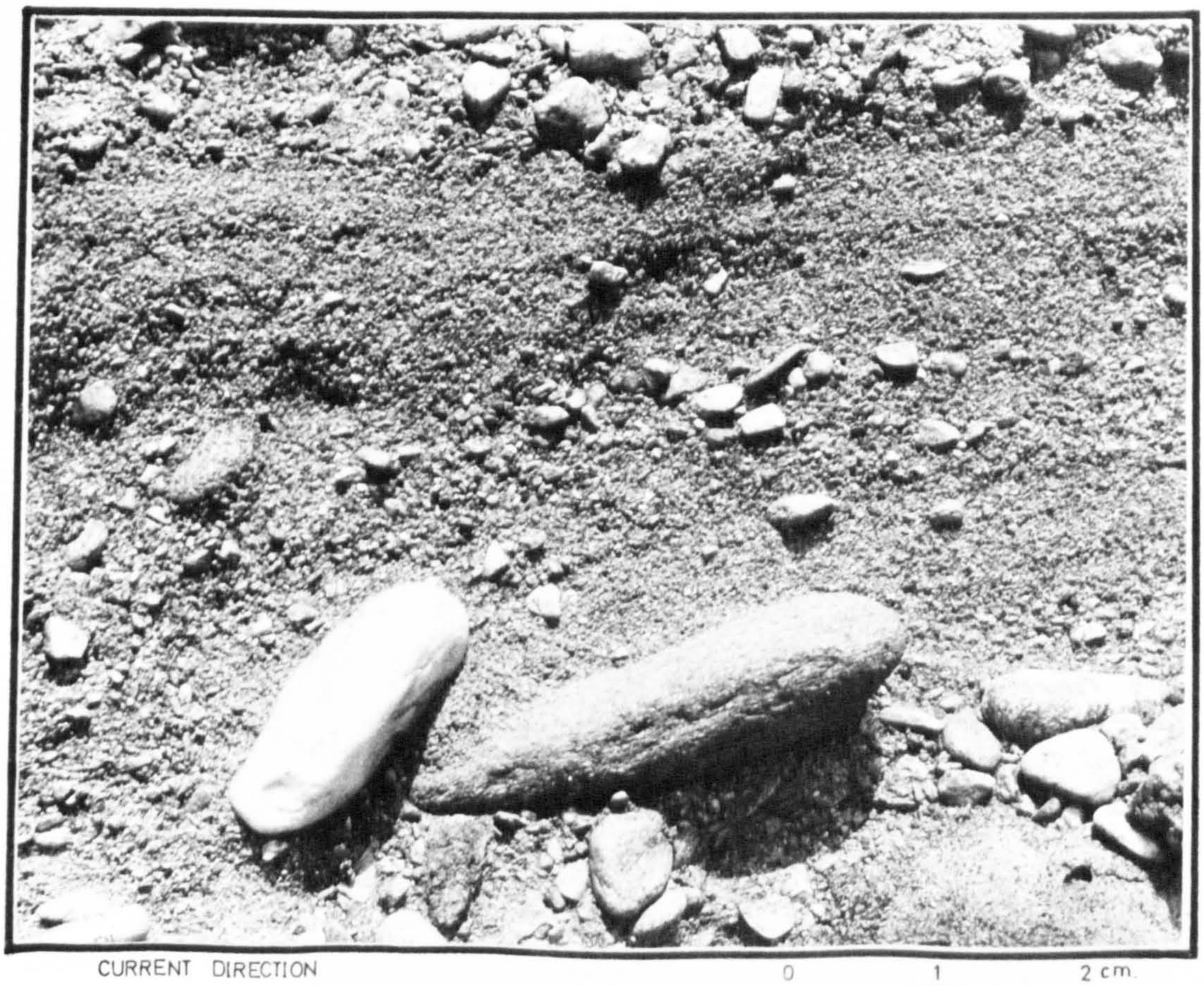


Figure 5.34
Isolate imbrication of coarse clasts, North Esk gravel pit



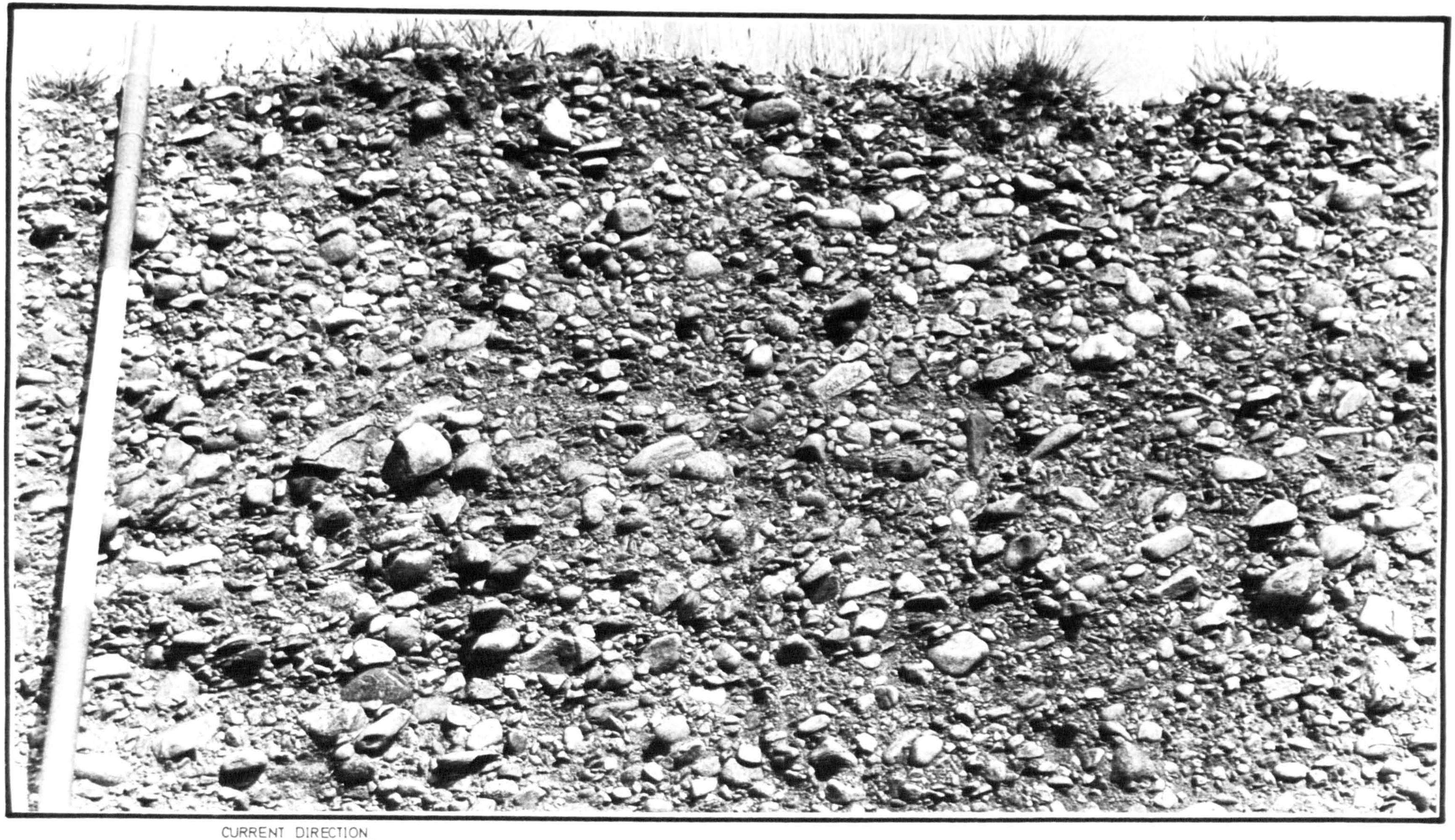


Figure 5.35 Contact imbrication of coarse clasts in paleo-outwash deposits, North Esk gravel pit

5.6.2.4. 'Antidune' sedimentation

Few 'antidune' structures were found in any of the outwash deposits, although it is likely that breaking waves, which produce a pattern of sediment waves on the stream bed, would have developed during periods of high stream velocity and Froude number. The bed-waves or 'antidunes' form where the channel slope is locally steep, and are built up from plane-bed particles into trains of 'dunes' which may move upstream, downstream or remain stationary (e.g. Menard, 1950). A backset series of inclined bedding develops (Fahnestock, 1963; Harms and Fahnestock, 1965; Simons et al., 1965; Jopling, 1969) which is inclined upstream at a relatively low angle (less than about 10°, Harms and Fahnestock, 1965) (e.g. Figure 5.36, page 306). The coarseness of the outwash deposits during high flow régime conditions has probably precluded the development of clearly-bedded antidune structures.

5.6.3. CONCLUSION

Both the Bossons and Strathmore outwash deposits comprise a complex series of overlapping channel and bar facies. Channels are more localized at Bossons, but the wider the channel the greater the variety of bed relief, bedforms and particle sizes across the channel floor. The bar facies, by contrast, is relatively more extensive at Bossons than in Strathmore. The bar deposits in each area are, nevertheless, composed of planar or crudely horizontal bedding deposited during periods of upper flow régime conditions.

Channel deposits, rather than bar or bar-top deposits, are more widely preserved in the Strathmore outwash deposits, especially in the downstream reaches. Examples from the South Esk gravel pit indicate mainly low flow régime sedimentation conditions, with rapid and extensive variations in channel position as stream aggradation progressed (Figure 5.24, page 292 ; Figure 5.29, page 298 ; Figure 5.30, page 299). Coarse, massive, imbricated gravels are, however, predominant in the upstream reaches. Structures formed of finer sediments have only been partiall

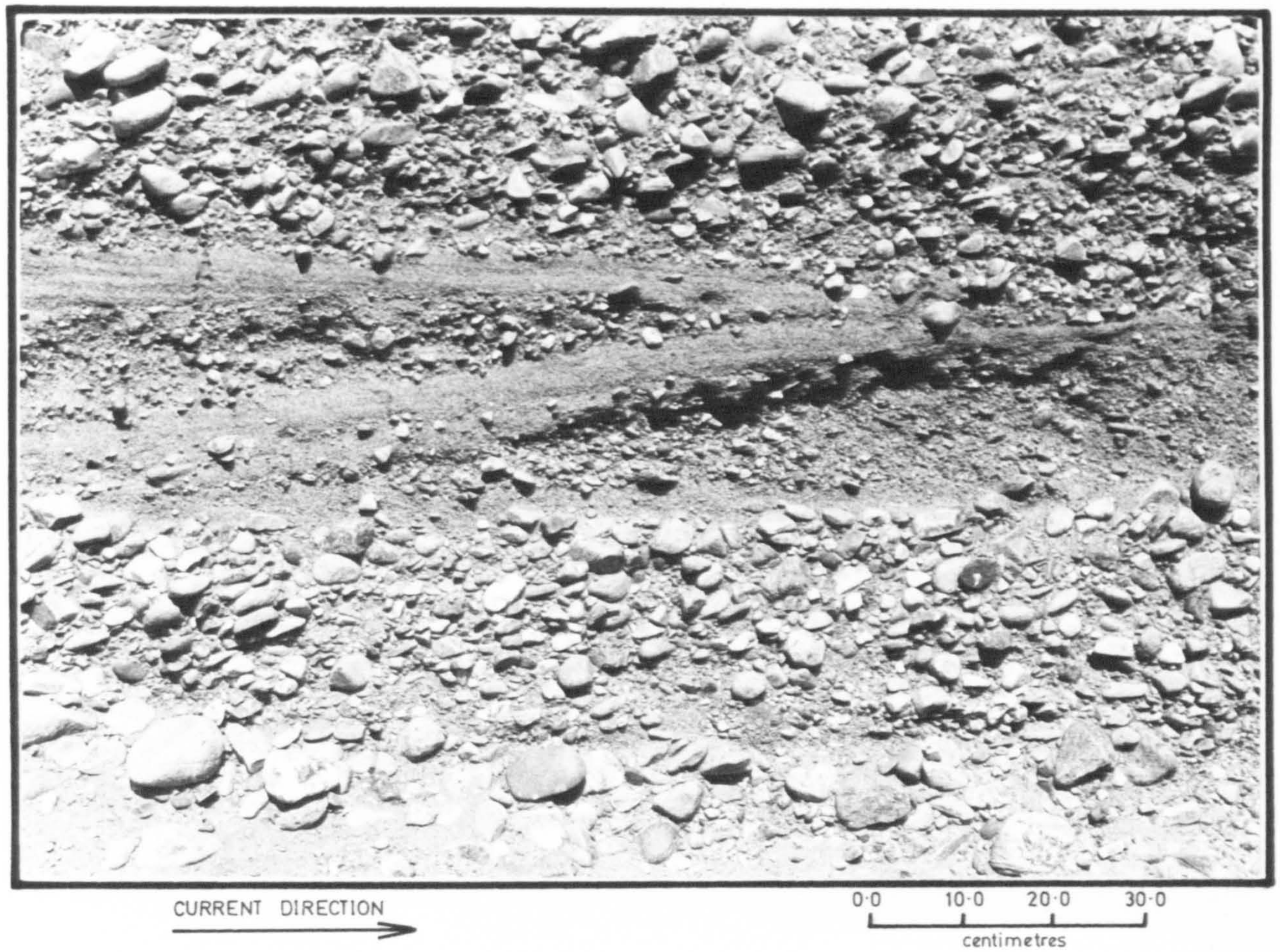


Figure 5.36 Possible antidune structure in paleo-outwash deposits, North Esk gravel pit

preserved in areas where coarse clast sedimentation has been most extensive.

The sedimentary structures deposited under similar flow régimes in the two series of deposits are directly comparable. The bedforms can be identified in plan in the present outwash environment and in section in the paleo-deposits, and because of the close relationships between flow régime, particle size and sedimentary structure, many aspects of flow conditions can be determined in the ancient outwash deposits.